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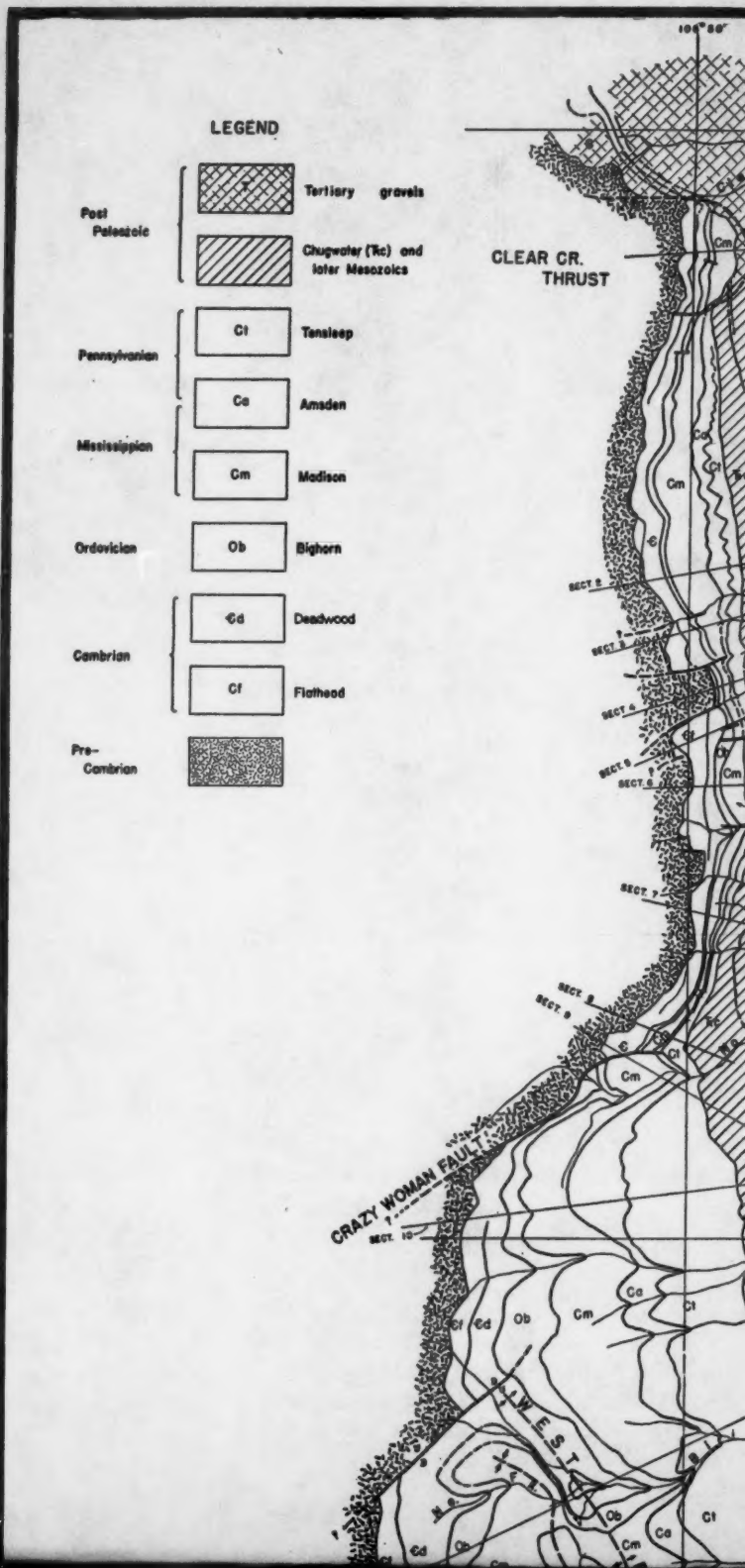
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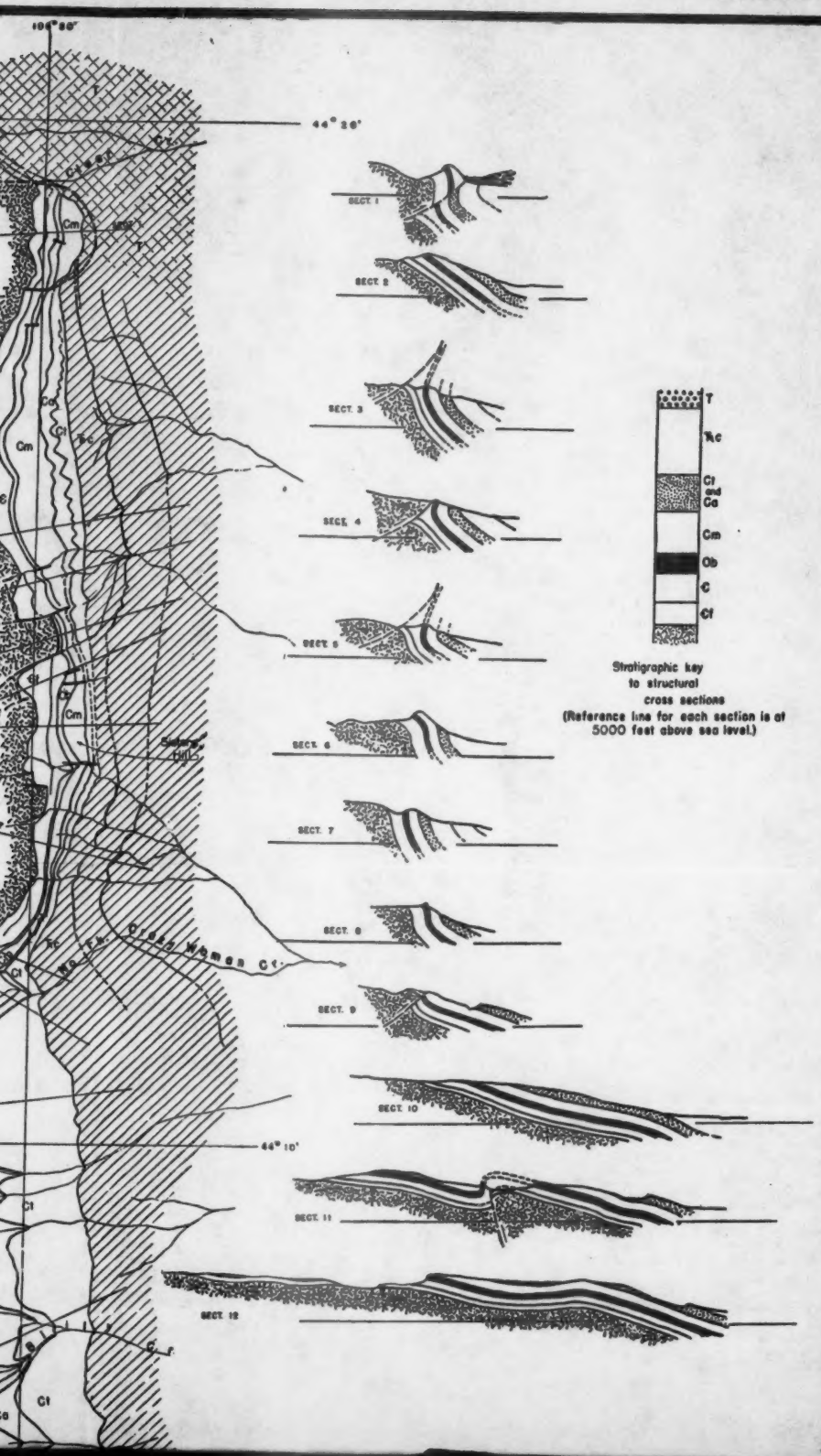
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Stratigraphic key to structural cross sections
(Reference line for each section is at 5000 feet above sea level.)

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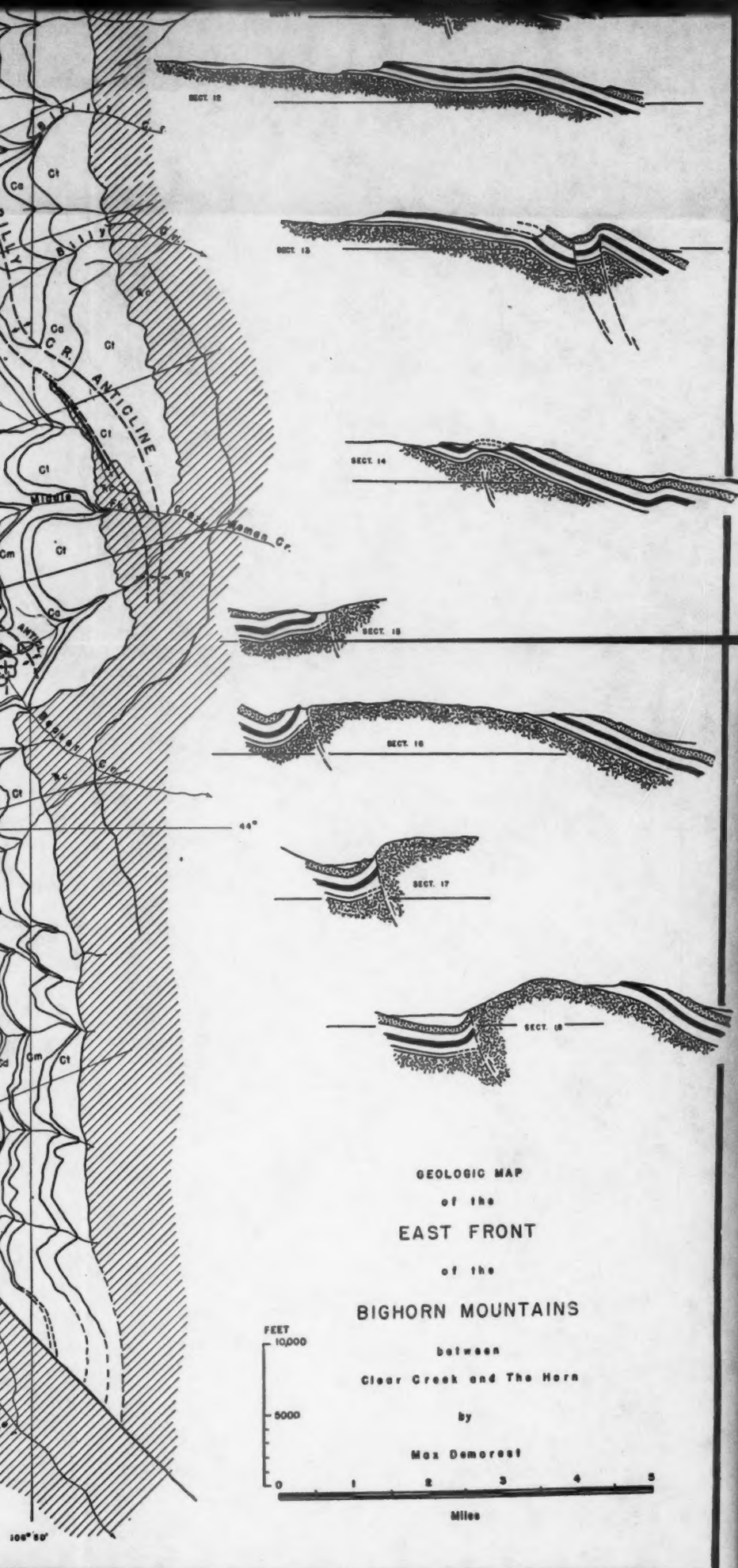

EAST FRONT OF BIGHORN MOUNTAINS
en Clear Creek and The Horn
outhern half of central segment)

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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CRITICAL STRUCTURAL FEATURES OF THE BIGHORN MOUNTAINS, WYOMING

BY MAX DEMOREST

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ABSTRACT

Structures studied in the Bighorn Mountains, principally along its eastern side, lead to a hypothesis of the structural history of the range. Apparantly the Laramide uplift took place during two distinct episodes, here referred to as the primary and secondary deformations. In the primary deformation uplift by compression caused the central part of the range to be arched asymmetrically with the steeper limb on the east. The northern and southern parts were asymmetrically arched to a lesser extent, with steeper limbs to the west. During the secondary episode of deformation numerous relatively small thrust structures were superimposed on the primary

structure. Evidence that the two episodes were distinct is found in the stratigraphy of early Tertiary(?) gravels deposited along the eastern front of the range.

The existence of an "ancestral" Bighorn range during the pre-Laramide history of the area is indicated by the presence of basal Chugwater stream gravels containing pebbles derived from Cambrian formations.

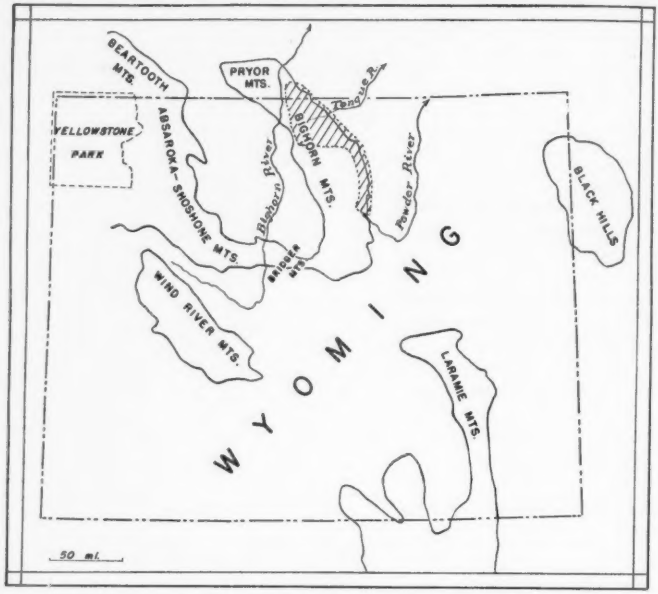

FIGURE 1.—*Index map of Wyoming*
Showing the area studied.

INTRODUCTION

LOCATION AND EXTENT OF THE AREA

The Bighorn Mountains form an arcuate range, convex eastward, lying between the Bighorn Basin on the west and the Powder River Basin on the east (Fig. 1). The length of the range is 120 miles, extending from the Pryor Mountains on the north to the Bridger Mountains on the south. The average width is between 25 and 30 miles. Structurally the range is a relatively simple asymmetrical anticline. The asymmetry, however, is not constant, for the range is divided into three segments the northern and southern of which are steeper on the west, whereas the central one is steeper on the east (Fig. 2). The boundary between the central and southern segments is sharply marked by the long Tensleep fault (Wilson,

1938) which strikes east-west. Between the central and northern segments the boundary, which lies along Tongue River, is less apparent.

The central segment is broader and higher than the others in both a structural and topographic sense. The end segments rise to elevations between 9000 and 10,000 feet (3000 to 5000 feet above the surrounding plains) with Paleozoic formations extending across much of the crest of the range. In contrast, the pre-Cambrian core of the central segment rises above 13,000 feet, and the Paleozoic formations are found only in hogbacks along its flank where they are truncated by a subsummit erosion surface at elevations between 7000 and 8000 feet (Figs. 2, 3).

Most of the present study was made along the eastern flank of the central segment of the range. Work was largely confined to the area in which Paleozoic and early Mesozoic formations outcrop in the hogback structures, yet numerous observations were made in the interior of the range in attempts to trace various faults across the pre-Cambrian core.

Detailed mapping was done in only the southern half of the area (Pl. 1) where structures shown on previous maps are not detailed enough for present purposes. This mapping was done by plane table with horizontal control taken from United States Geological Survey topographic maps. The inaccuracy and, in many places, the absence of township and section markers made the usual method of horizontal control impossible. In the northern half of the area all except a very few minor structures are well shown on existing maps (Darton, 1906a; 1906b; 1906c), and for that reason additional mapping was not necessary. (Plate 4 is compiled with only slight modification from Darton's maps.)

ACKNOWLEDGMENTS

Field work was done during three summers. That carried on in 1936 and 1937 was financially supported by the Geology Department of Princeton University and the Yellowstone-Beartooth Research Association, and the results were presented as a doctorate dissertation at Princeton. Further work in the summer of 1938 was supported by the Geological Society of America and the Yellowstone-Beartooth Research Association. The writer is also indebted to the University of Cincinnati and the American Geographical Society for the loan of instruments.

Field work was greatly aided by the kind assistance of people living in the area. Particular thanks are due Mr. and Mrs. Goelet Gallatin of Bighorn and Mr. and Mrs. Ralph Roe and the Hackert family of Klondike.

Professor W. T. Thom, Jr., of Princeton University made many helpful criticisms and suggestions during the progress of the work and thanks are also due him for many other kindnesses.

TABLE 1.—*Stratigraphic section*

Age	Formation		Lithologic characteristics	Thickness (Feet)
TERTIARY	KINGSBURY		Fine to very coarse gravel. Pebbles derived from pre-Cambrian and Paleozoic formations except basal part which lacks those from the pre-Cambrian.	0 to 2000
TRIASSIC	CHUGWATER Ꞧ		Red calcareous shale and sandstone with abundant gypsum.	730 to 1200 (Thins toward south)
PENNSYLVANIAN	TENSLEEP Ct		Gray, buff, and pink massive calcareous sandstone alternating with fine-grained buff clastic dolomite.	550
	AMSDEN Ca		Buff and pinkish fine-grained clastic dolomite with much chert. Pink and buff sandstone and red shale at base.	
MISSISSIPPIAN	MADISON Cm		Gray and white fine-grained clastic limestone. Occasionally cross-bedded. Some oölites. Highly fractured and recemented.	625
ORDOVICIAN	BIGHORN Ob		Buff and gray clastic dolomite and limestone. Some cross-bedding. Lower 50 feet red, buff, and white cross-bedded and massive coarse-grained sandstone.	275 to 285
CAMBRIAN	DEADWOOD Ꞓd	UPPER	Reddish irregular, thin-bedded calcareous sandstone, clastic limestone and shale. Some flat-pebble conglomerate.	425 to 450
		LOWER	Drab flat-pebble conglomerate and shale.	
	FLATHEAD Ꞓf		Brown to deep-red sandstone and quartzite. Conglomerate at base.	270 to 365 (Thins toward south)
PRE-CAMBRIAN			Complex metamorphic and intrusive formations.	

The writer is also indebted to his wife and to the following undergraduate and graduate students who at various times gave assistance in the field: M. G. Frey, Warner S. Goshorn, D. G. Herring, Jr., Robert W. Johnson, L. R. Kirk, and Vincent McKelvey.

STRATIGRAPHY

With two exceptions the stratigraphy of the area has been studied by the writer only insofar as it bears on the geologic structures. The generally accepted definitions of Darton (1906a) have been followed, and formations were identified and described entirely on the basis of lithology. This is possible because there is little lateral variation in the character of the formations, and because extensive exposures make it possible to trace each formation throughout the entire Bighorn area. The only deviation from Darton's definitions has been a separation of the basal sandstone and quartzite member from the Deadwood formation. This follows the usage of Wilson (1938), who calls this member the Flathead on the assumption that it is correlative with the basal sandstone and quartzite of other parts of the Middle Rocky Mountains. For purposes of description the remainder of the Deadwood is also subdivided, distinctive lower and upper parts being recognized.

The stratigraphic section for the area includes Cambrian to Triassic shales, limestones, and sandstones, and early Tertiary (?) gravels. The sequence and most of the important information concerning these formations is summarized in Table 1.

STRUCTURE

GENERAL STRUCTURAL RELATIONSHIPS

The critical structural features studied indicate that (1) the present Bighorn Range was uplifted during at least two distinct episodes of deformation and (2) it was foreshadowed by an "ancestral" range in late Paleozoic or early Mesozoic time. Figure 2 shows the principal structural relationships on which the first of these statements is based. The important facts are:

(1) The asymmetry of the central segment with its steeper limb on the east is in contrast with the westward steepening of the northern and southern segments.

(2) Of the many smaller thrust structures on the eastern side of the range only those along the middle part of the central segment are overthrust toward the east (Pl. 1, secs. 1, 3, 4, 5; Pl. 4, sec. 3); those of the end parts of the central segment (between the Piney Creek thrust and Tongue River at the north and between the Crazy Woman fault and The Horn at the south) oppose the larger structure of that segment but are in conformity with the larger structures of the northern and southern segments (Fig. 2; Pl. 1, secs. 11, 13-18; Pl. 4, sec. 2).

These relationships are interpreted as showing that the larger structures are of an earlier origin. It is probable that the first rise of the present

mountains resulted from gentle uparching and incipient overthrusting wherein the central segment was steepened on the east while the northern and southern segments, overthrust in the opposite direction, were steep-

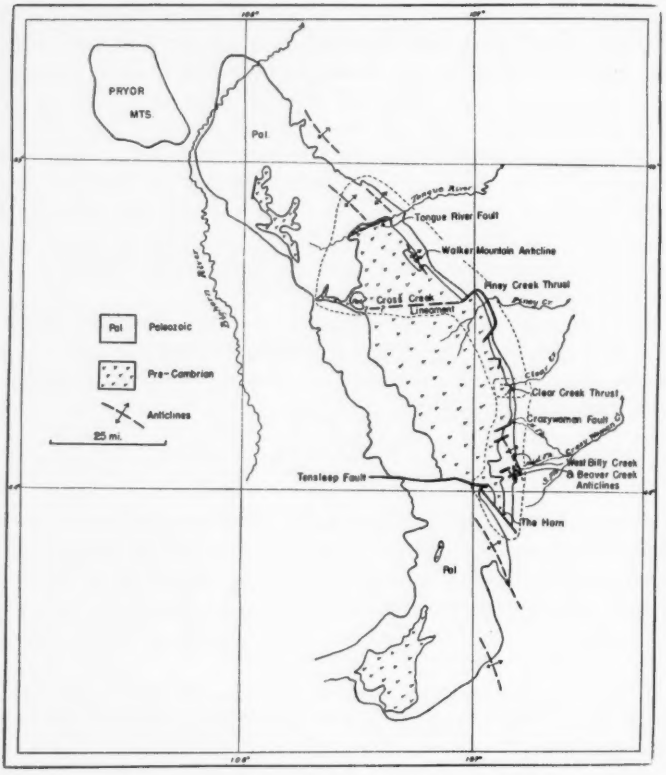

FIGURE 2.—*Map showing the principal structural elements of the Bighorn Mountains*

ened on the west. Such opposed thrusting produced the well-defined Tensleep fault which separates the southern and central segments and the less conspicuous faulting along Tongue River between the northern and central segments. Later deformation produced the smaller secondary structures in which only the middle part of the central segment continued to be overthrust eastward. At this time the Crazy Woman fault and the

fault at the northern end of the Piney Creek thrust mass were the tears between the areas of opposed thrusting.

It must be emphasized that this interpretation is only a hypothesis, for its validity depends upon certain assumptions which can be neither proved nor disproved. Yet these assumptions are of value in that they link the various structures of the area together in a unified and logical (though

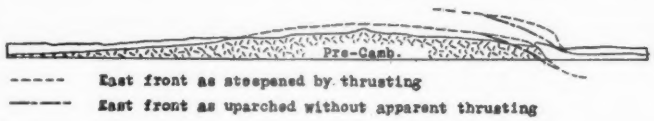

FIGURE 3.—*Topographic and structural cross section*
Central segment of the Bighorn Mountains.

hypothetical) interpretation. If the conclusions drawn are of regional significance future work in the Bighorns or in other parts of the Rocky Mountain area may help to either substantiate or disprove them.

The assumptions will be pointed out as they arise in the detailed discussion of the structures.

DETAILED STRUCTURAL DESCRIPTIONS

Middle division.—Within the middle division [1] there are only two localities of relatively gentle dips. These are at places where the secondary overthrusting was least effective, and therefore they probably show the character of this eastern front of the mountain previous to the secondary oversteepening. One of these localities is shown in Section 2 of Plate 1; the other, not shown in cross section, is just south of the Piney Creek thrust mass (Pl. 4). Figure 3, a generalized section across the range, shows that the oversteepened part of the eastern flank is localized near the edge of the uplift and that, except for the secondary thrusting, the eastern limb would be only slightly steeper than the western. This is also shown in Sections 1, 6, and 8 (Pl. 1) and in Figure 1 of Plate 2 in which it is seen that the tops of the higher hogbacks dip more gently, curving into the flatter slopes that continue toward the crest of the range.

The northernmost and largest eastward thrust structure is at Piney Creek (Pl. 4, sec. 3); the tear along its northern end shows a stratigraphic displacement of 3 miles in which about 2½ miles of pre-Cambrian rock is exposed. The position of the fault plane underlying the thrust mass is indeterminable, for the fault line along the front of the mass is covered by early Tertiary (?) gravels (Pl. 2, fig. 2). The hypothetical fault plane

[1] The natural subdivision of the eastern front of the central segment is made use of for ease of description. The middle part, where the secondary thrusting was to the eastward, is here called the middle division. The other two parts are called the northern division and the southern division.

shown in the structure section (Pl. 4, sec. 3), is based on analogy with other thrusts in the area. It presents the simplest form in which the thrusting may have occurred though it is known that more complex faulting, producing the same general effect, took place at the southern end of the Piney Creek thrust mass (Bucher *et al.*, 1934, sketch page 169).

The gravels covering the fault plane are of especial interest, for they definitely show that the primary uplift of the range preceded the secondary thrusting at this place (and presumably elsewhere along the mountain front).[2] The following facts are cited as evidence: (1) The gravels lie uncomformably over the steeply dipping Cretaceous beds which were folded with the Paleozoic formations in the primary uparching of the range and later eroded to the present level of the plains. (2) The lower gravels contain abundant pebbles derived from the Paleozoic formations. Either of these facts is sufficient to show that deposition of the gravels postdated the uplift of the range. In addition the first one shows that there was enough time between the uplift and the gravel deposition to allow the Mesozoic formations to be reduced to the present level of the plains. Added to this is the fact that the lower gravels at the northern end of the thrust mass contain no pebbles which could have originated from pre-Cambrian rocks, thus showing that the Piney Creek faulting with its attendant exposure of pre-Cambrian rocks must have occurred after the deposition of the lower gravels and therefore considerably after the primary uplift of the range.

Unfortunately the dating of these gravels is uncertain. Darton (1906a; 1906b; 1906c) called them the Kingsbury conglomerate and gave their age as uppermost Cretaceous. The more recent work of Knowlton (1909) and Wegemann (1917), however, suggests that they may be basal Wasatch.

Faults at Clear Creek and at Sections 3, 4 and 5 (Pl. 1) offer the best evidence for the ramplike shape ascribed to the thrust planes. This evidence is not entirely conclusive, for none of the thrust planes is well exposed, yet the geometry of the folded Paleozoic formations and the topography of the present hogbacks make it clear that the faulting was directed both eastward and upward along curved surfaces.[3] The fact of eastward thrusting is obvious, while the upward component of movement is shown at Sections 3 and 4 (Pl. 1) where the thrust masses rose over the

[2] It is assumed that all the secondary thrusts are contemporaneous. This is suggested, but not proved, by the fact that all of them, whether overthrust eastward or westward, strike in approximately the same direction as if developed by a single uniformly directed compressive force.

[3] The author does not intend to infer that the fault planes were smooth surfaces of displacement. It is more likely that the displacement involved zones of distributed fracturing and minor faulting. However, since all the thrust masses appear to have moved as discrete blocks these zones may be thought of as approximating surfaces, which were the boundaries of the blocks.

hard Flathead quartzite and sandstone without producing drag, yet caused overturning of the Bighorn and higher Paleozoic formations. A plastic deformation of the Deadwood shales and limestone seems to have occurred to fill the gaping space left between the undragged and overturned formations.

Southern division.—The maps (Fig. 2; Pl. 1) show the three significant secondary structures of the southern division. Each of them is essentially an asymmetrical anticline produced by overthrusting toward the west-southwest. Differences within and between the structures are due principally to variations in the amount of thrusting.

The West Billy Creek structure when viewed in its entirety is regarded as a long thrust block striking obliquely across the older primary mountain front at an angle of about 20 degrees. Thus the southern end of the block departs from the older mountain front to form a long spur, whereas its northern part crosses the front at higher and higher elevations to the point where it is terminated on a tear fault. The amount of thrusting, except in the northern part of the structure, was not sufficient to cause faulting at the surface. As a result it exists as an anticline, the steeper western limb of which represents the "scarp" of the thrust block. (*See* Sections 11, 12, and 13 of Plate 1.)

Associated with this structure and parallel to it is the shorter fault which broke through to the surface in the synclinal axis just west of the anticline (Pl. 1, sec. 13). It is probably a subsidiary thrust associated with the larger structure. Proof of this is lacking. However, it is the only logical interpretation consistent with the general structural pattern of the area.

Detailed study of the larger structure shows that the thrust block was segmented, its different parts showing different characteristics. The northern part (Pl. 1, sec. 11) where the fault is apparent has been dissected deeply by the headwaters of North Billy Creek. Perhaps the faulting is apparent only as a result of the dissection, yet the high degree of asymmetry here shows that the thrusting was more effective than in the middle part of the anticline which lies immediately to the south. (*Compare* Sections 11 and 12 of Plate 1.) The transition between the two parts is very abrupt suggesting that a cross fault may separate them. This is further suggested by the fact that a small branch of Billy Creek has cut deeply across the structure at this place. However, any positive structural evidence that may have existed for such a fault has been destroyed by weathering and erosion in the creek valley.

Between the middle and southern parts of the anticline there is another though less marked transition, the southern part being more asym-

metrical, topographically lower, and slightly offset to the east. (*Compare* Sections 12 and 13 of Plate 1.) The southern part also differs from the others in that its axis plunges southward. Where the hard Tensleep sandstone covers the axis, it forms an excellent anticlinal ridge (Pl. 2, fig. 3). Farther south, beyond the point where the Tensleep sandstone plunges beneath the Mesozoic formations of the plains, the anticline has little topographic expression though it continues for at least a mile. Its exact termination is indefinite, but, for reasons given below, it may be suspected to end along a line marking an eastward prolongation of the Tensleep fault.

The short Beaver Creek structure is similar to the middle part of the West Billy Creek anticline in that it is nearly symmetrical, probably a result of less effective thrusting. The northern termination of this structure is thought to be a tear along the Middle Fork Crazy Woman Creek. Erosion of the creek canyon has removed any direct evidence of tearing, but the fact that the sedimentary formations on the northern side of the canyon are not folded is sufficient to show that the structure ends abruptly (Pl. 3, fig. 1). The southern termination of the structure is also abrupt. Whereas the West Billy Creek and Horn structures continue into the plains as long spurs this one produces only a jog in the mountain front, giving way within a few yards to a steep dip slope that carries the Paleozoic formations beneath the plains. It is notable that the line on which this ends is also the prolongation of the Tensleep fault.

This last fact is of particular significance, for the northernmost of the tears at the northern end of The Horn structure is also along the same Tensleep line. Since the Tensleep fault presumably originated with the primary uplift of the range, these relationships are interpreted as an indication that it acted as a line of weakness along which the northern end of The Horn and the southern end of the Beaver Creek structures were localized at the time of the secondary uplift. It is also probable that the southern end of the West Billy Creek structure may lie along this same line.

Unfortunately the eastward prolongation of the Tensleep fault cannot be traced farther than the tear at the northwest of The Horn. Beyond that point any definite trace of the fault is lost in the pre-Cambrian rocks. However, its connection with the northernmost of The Horn tears is very definite, for the two faults can be traced as a single line along which the displacement is reversed at the point where The Horn thrusting became effective in re-elevating the southern side.

Southward from the line of the Tensleep fault the eastern side of The Horn thrust block (the dip slope) constitutes a continuation of the front

FIGURE 1. NORTHERN END OF THE CLEAR CREEK THRUST MASS
Taken from north side of Clear Creek Canyon showing high angle of dip at front of thrust mass and low angle of dip at top.

FIGURE 2. NORTHEASTERN EXTREMITY OF THE PINEY CREEK THRUST MASS
Facing south. (1) Pre-Cambrian of the thrust mass; (2) Paleozoic hogback at the front of the thrust mass; (3) Tertiary (?) gravels deposited against front of thrust mass.

FIGURE 3. WEST BILLY CREEK ANTICLINE AND SYNCLINE
Facing east-southeast. Ridge of Plunging anticline (left), syncline (center), eastward dip slope of the Beaver Creek anticline (right).

CLEAR CREEK AND PINEY CREEK THRUST MASSES, AND WEST BILLY CREEK ANTICLINE

FIGURE 1. BREACHED NORTHERN END OF THE BEAVER CREEK ANTICLINE
Facing north-northwest. In distance flatlying sediments of the northern side of the Middle Fork Crazy Woman Creek.

FIGURE 2. HORN FAULT ESCARPMENT
Facing south-southeast. Dip slope of southern segment of range (right), here making western limb of syncline formed by The Horn thrusting.

FIGURE 3. WALKER MOUNTAIN STRUCTURE
Facing east, looking across breached Walker Mountain anticline. Light-colored stratum dipping to left is Madison limestone of Walker Mountain dip slope. The fault lies along the strike of the lower edge of this stratum. Scarp slopes of the main mountain front hogbacks in background.

VIEWS OF THE BEAVER CREEK, THE HORN, AND WALKER MOUNTAIN STRUCTURES

of the central segment of the range. On its southwestern side the block is bounded by a steep escarpment (Pl. 3, fig. 2) which marks the outcrop of the thrust fault. At its southern end the amount of thrusting decreases to a point where the hard rocks of the pre-Cambrian and Paleozoic formations are not brought above the level of the plains. There the topographic expression of the structure ends, but the faulting continues into the plains, finally giving way to an unfaulted anticline.

The wide embayment of the plains which lies west of The Horn owes its existence to the fact that the front of the southern segment of the range lies farther west than the front of the central segment, the two fronts having been offset by the Tensleep fault in the earlier deformation. Farther south along the front of the southern segment is another and larger anticline (Fig. 2) which is also considered to be a thrust structure, and, though proof is lacking, it is believed to be related in origin to the secondary structures of the central segment.

Northern division.—The two secondary structures of the northern division (Pl. 4) are unlike those of the southern division in that they approximately parallel the mountain front; this is due to a difference in the strike of the mountain front, rather than to a difference in the strike of the secondary structures (Fig. 2).

The smaller of the two structures of the northern division is an asymmetrical anticline. At its northwestern end, where the axis disappears in the pre-Cambrian formations, a small fault along the steeper western limb indicates the thrust condition responsible for the structure.

The larger Walker Mountain structure is even more obviously a result of thrusting, for it shows two parallel high-angle faults dipping northeastward. The smaller thrust, closer to the mountain front, grades into an anticline at its northwestern end where it crosses Wolf Creek. Beyond there the anticline gradually dies out within about 2 miles. No tear fault was found. Southwestward from Wolf Creek the fault can be traced intermittently for approximately 5 miles where it ends by grading into a broad fold of low amplitude. This is opposite the southeastern end of Walker Mountain.

As seen on the map (Pl. 4), the larger thrust lies about 1 to 1½ miles farther in from the mountain front with Walker Mountain rising from its southwestern side as a second hogback of Paleozoic formations (Pl. 3, fig. 3). Throughout most of its length the upthrown side of the fault is of pre-Cambrian rock, exposed as a result of deep erosion by a branch of Wolf Creek (Pl. 4, sec. 2). The downthrown side is the Madison limestone of the Walker Mountain dip slope. Had the thrusting produced only a folding of the Paleozoic formations without breaking

through to the surface, this dip slope would constitute the southwestern limb of a syncline. In the southern division where the secondary structures lie oblique to the mountain front such synclines are simply the depression left between the upraised thrust block on one side and the older mountain front on the other. However, here at Walker Mountain this explanation will not hold, for the southwestern side of the syncline does not correspond with anything that might be considered as the older mountain front. It is more likely that the older front is coincident with the present one and that the thrusting has modified it only by producing steeper dips.

This being the case the Walker Mountain dip slope is best interpreted as a remnant of the gently dipping formations which once extended from the top of the more steeply dipping mountain front to the crest of the range. The present increased dip of the formation is probably due to tilting produced during thrusting. Such tilting is presumably a result of a possible though indeterminable thrust fault lying parallel to and southwest of Walker Mountain. If such a fault exists Walker Mountain is a fault block which was tilted by overthrusting. Evidence in support of this possibility is found at the ends of Walker Mountain where the Paleozoic formations are bent upward producing local dip slopes which strike at an angle to the main structure. Such bending may be due to either drag along terminal tear faults, or to flexing, or to both. In any case it demonstrates that the Walker Mountain block has moved relative to the areas lying off its ends.

Tongue River fault.—The selection of Tongue River as the boundary between the central and northern segments of the range is based mostly on the fact that the exposed pre-Cambrian core of the range ends there (Fig. 2; Pl. 4). Though no good evidence of any single clear-fault can be found, there is no doubt that a considerable degree of faulting, perhaps on multiple planes, has occurred within a fairly narrow zone along the river in the eastern half of the range. This is demonstrated by numerous slickensided outcrops where deformation has been intense enough to produce "rolled" surfaces in the hard rocks, and also by the attitude of the Paleozoic formations on the northern side of the river (Pl. 4, sec. 1). There the formations are sharply upturned, either as a result of drag along a large fault, or perhaps as a result of flexing accompanied by numerous minor faults. In either case a major structure is present, at least in the eastern half of the range.

The western side of the range, beyond the exposed pre-Cambrian core, offers an entirely different aspect. In that area there is no evidence of faulting. The relatively flat-lying Paleozoic formations which cover the

crest of the northern segment extend without apparent displacement into the central segment, continuing uninterrupted for 10 miles to the vicinity of Shell Creek (Fig. 2; Pl. 4). In this connection it must be noted that the steeply dipping western flank of the northern segment also continues without interruption to the vicinity of Shell Creek. It is only south of there that the gently dipping western flank, typical of the central segment, is to be found. With these relationships in mind a person studying the western side of the range would logically place the boundary between the northern and central segments near Shell Creek, rather than along Tongue River as the present author has done. The point of view depends upon the side from which one approaches the range. This fact brings added significance to another lineament that extends at least part way across the range—namely, the Cross Creek lineament.

Cross Creek lineament—This lineament, like the Tongue River fault, seems to be a major structure transverse to the range. Also, like the Tongue River fault, it is difficult to trace with any degree of accuracy. As shown on the maps it appears to be a westward continuation of the tear fault that marks the northern end of the Piney Creek thrust mass. The author flew over the area and was able to see the lineament as a clearly defined feature trending across the range as far as the head of Big Goose Creek (Pl. 4). When studied from the ground, however, it was more difficult to determine. Along the east-west section of Cross Creek the only evidence of faulting, apart from the unusual straightness of the canyon, was found in fairly numerous slickensided pebbles and boulders. Unfortunately none of these could be found in place; yet most of them were angular, thus indicating a near-by origin. This evidence, however, is not necessarily indicative of faulting along a single line, for slickensided specimens are common in other parts of the pre-Cambrian core, both north and south of Cross Creek. Many of these originated along faults of small displacement which appear to be scattered throughout the interior of the range.

The evidence of major faulting along the head of Big Goose Creek is better, for there the opposite sides of the canyon show distinctly different kinds of foliated rocks. The possibility that the creek might mark a pre-metamorphic contact rather than a fault is eliminated by the fact that the attitude of the foliae is quite different on the two sides of the canyon.

Though these evidences for the lineament are not the best, they are nevertheless convincing, especially when supplemented with a view from the air. Unfortunately photographs from the air were not satisfactory.

The connection of the lineament with the change of structure near Shell Creek (Fig. 2), on the western side of the range, could be neither proved nor disproved. Evidence of the lineament ends at the head of Big Goose Creek, 4 miles from the main divide and 16 miles from the western foot of the range. Yet the suggestion of its continuation seems pertinent, for such a transverse line may logically be explained as a tear produced between eastward and westward thrust parts of the range at the time of the secondary deformation. Thus it would seem that the active boundary between the central and northern segments changed from the Tongue River fault to the Cross Creek lineament and that the secondary deformation was responsible for the highly steepened western front of the range north of Shell Creek.

"ANCESTRAL" BIGHORNS

Basal Chugwater conglomerates found at various places along the eastern front of the mountain contain pebbles derived from Cambrian rocks, thus indicating that an "ancestral" range was uplifted and deeply eroded in early Chugwater time. The position of this earlier uplift is not certain. However, is seems probable that it occupied the site of the present mountains. Confirmation may be sought in further studies of the Chugwater formation at other places along the mountain front. The age of the basal Chugwater conglomerates, and therefore of the "ancestral" uplift is also uncertain, though it is known to be either late Paleozoic or early Mesozoic. This approximately corresponds with the age of other "ancestral" ranges known elsewhere in the Rocky Mountain region (Ver Wiebe, 1930).

SUMMARY

The anticlinal Bighorn range is structurally divisible into three distinct segments—northern, central, and southern. The central one is highest and broadest, and asymmetrical with the steeper limb to the east. The other segments are also asymmetrical, but with the steeper limbs to the west. It is believed that this asymmetry expresses a slight overriding or incipient overthrusting toward the steeper limbs.

On the basis of secondary structures the eastern side of the central segment is also divisible into three divisions—northern, middle, and southern. These secondary structures are thrusts which, in the middle division, are overthrust toward the east-northeast, and in the northern and southern divisions overthrust toward the west-southwest. Thus the secondary structures of the northern and southern divisions of the segment are thrust in conformity with the primary and secondary thrusting of the northern and southern segments and in opposition to the

primary thrusting of the central segment. This opposition of the secondary structures to the primary structure of the central segment suggests that the two are not contemporaneous. Support for this suggestion is found in the Tertiary (?) gravels lying off the eastern front of the range, for the primary uplift and a considerable amount of erosion occurred there previous to the development of one of the secondary structures.

The boundary between the southern and central segments is along the well-defined Tensleep fault (Wilson, 1938) which strikes east-west across the entire range. The boundary between the central and northern segments is less pronounced, there being no single well-defined fault extending across the entire range, but there is a major zone of faulting or flexing along Tongue River at the northern end of the exposed pre-Cambrian core. As viewed from the eastern side of the range this appears to be the logical boundary between segments, and for purposes of discussion in the present paper it has been used as such. However, when viewed from the western side of the range this boundary is much less appropriate, for on that side there is no evidence of such a fault. The characteristic steep western limb of the northern segment continues without significant interruption to the vicinity of Shell Creek, 10 miles south of the line of the Tongue River fault. Thus, when seeing the range from the western side, one would place the boundary near Shell Creek. In this connection significance is found in the Cross Creek lineament which strikes east-west across the eastern half of the range and may very probably continue all the way across. Its eastern end marks the boundary between the middle and northern divisions of the central segment, and its western end, if continuous, would intersect the opposite side of the range near Shell Creek where the steep western limb of the northern segment gives way to the gentle western dip slope of the central segment.

Probably the Tongue River fault marks an early boundary between segments and was formed during the primary deformation. Later, when the secondary thrusting occurred, the boundary shifted to the Cross Creek lineament with the result that the minor structures of the northern division of the central segment were developed in opposition to the primary structure. At the same time the western side of the northern division became steepened in the same way as the western side of the northern segment.

From a regional point of view it is notable that the secondary structures of the Bighorns, all of which are subparallel in strike, are also parallel to minor structures in other ranges of the Middle and Southern Rocky Mountains. Striking examples are seen in the spurs that extend obliquely

from the Front Range of the Southern Rockies and in the Rattle Snake Mountain spur of the Absaroka Mountains. Broader regional studies may show that these other ranges were uplifted in two episodes of deformation and the parallelism of the many secondary structures may reveal a widespread compressive force of uniform direction.

Stratigraphic evidence is presented to show that the present Bighorn Mountains were preceded by an "ancestral" range during early Chugwater time (late Paleozic or early Mesozoic). It is notable that this also relates the Bighorn area to a broader regional phenomenon (Ver Wiebe, 1930).

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Yale University, New Haven, Connecticut.

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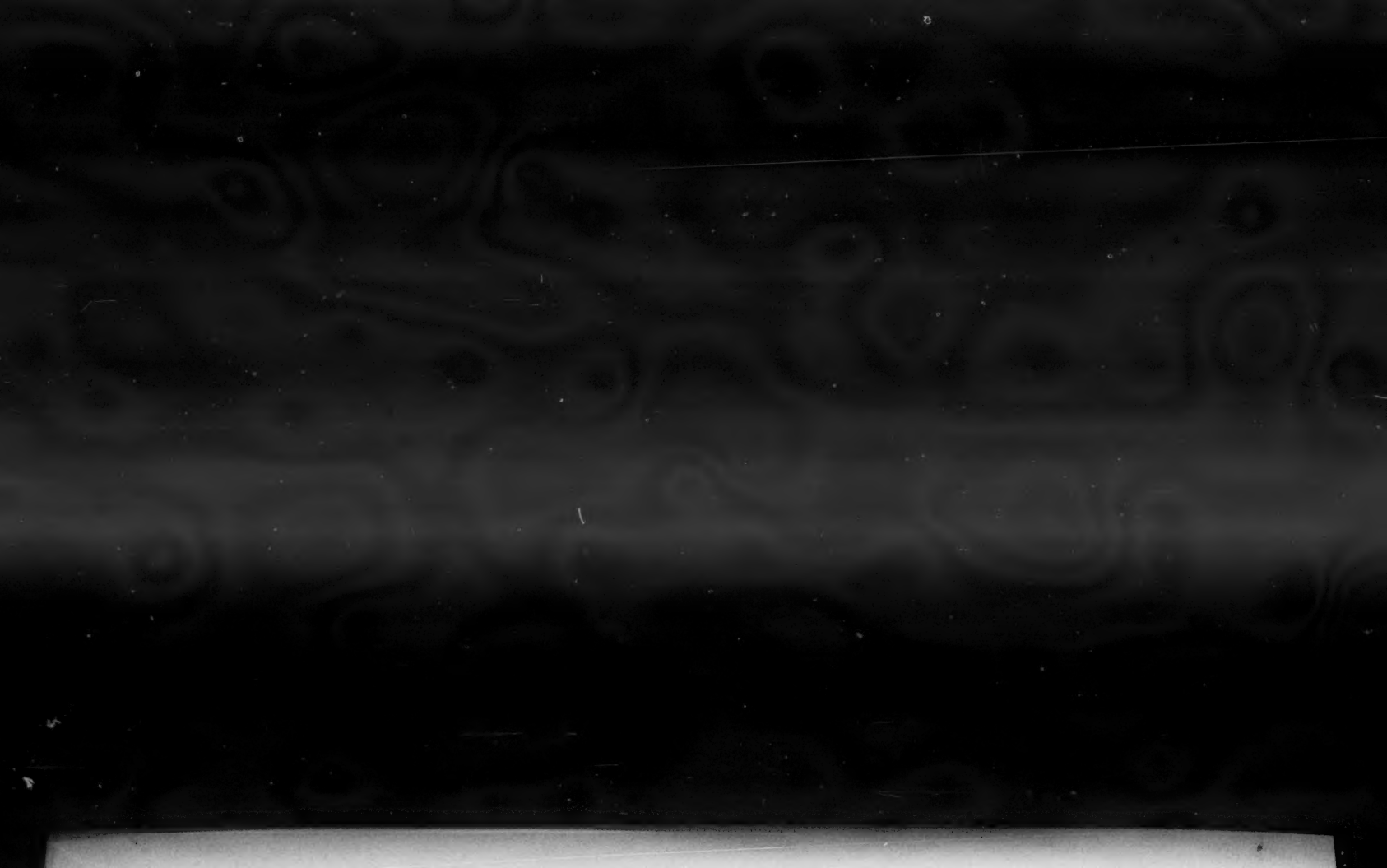

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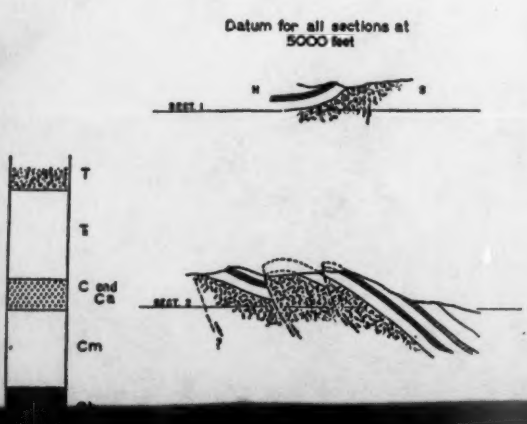
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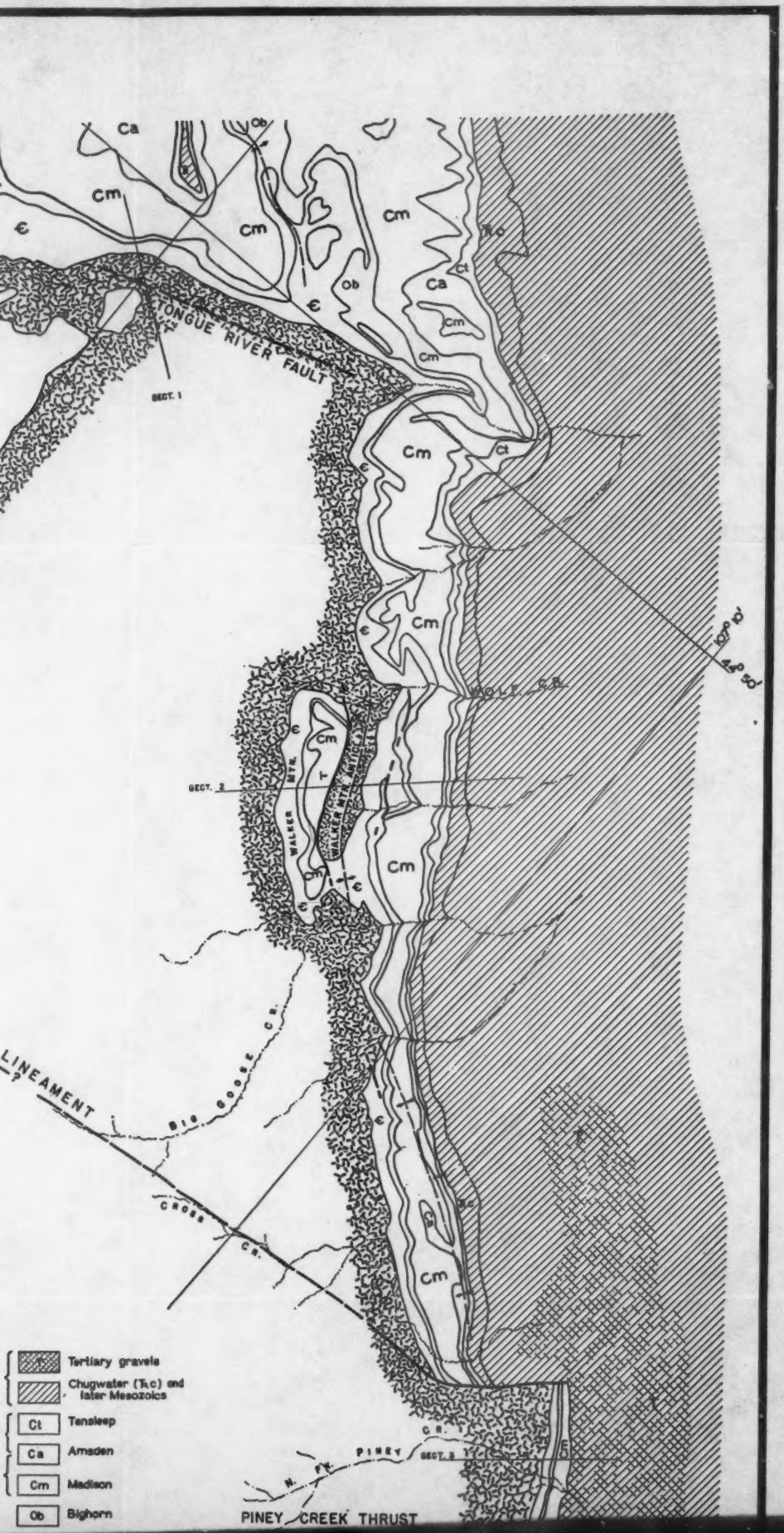
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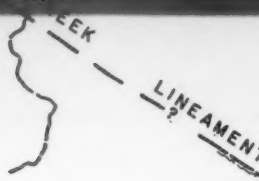

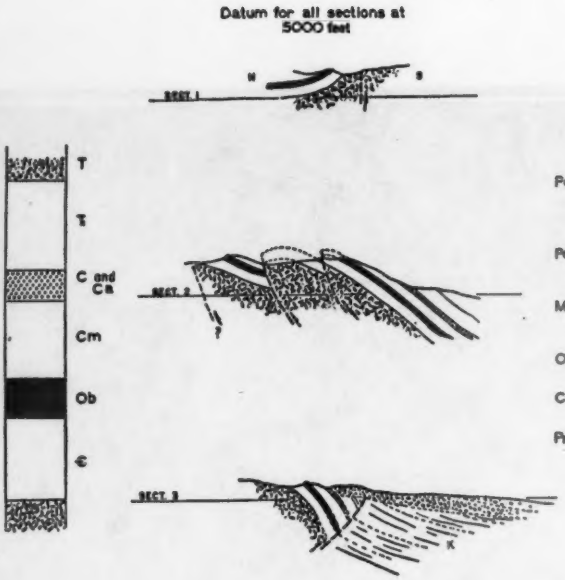

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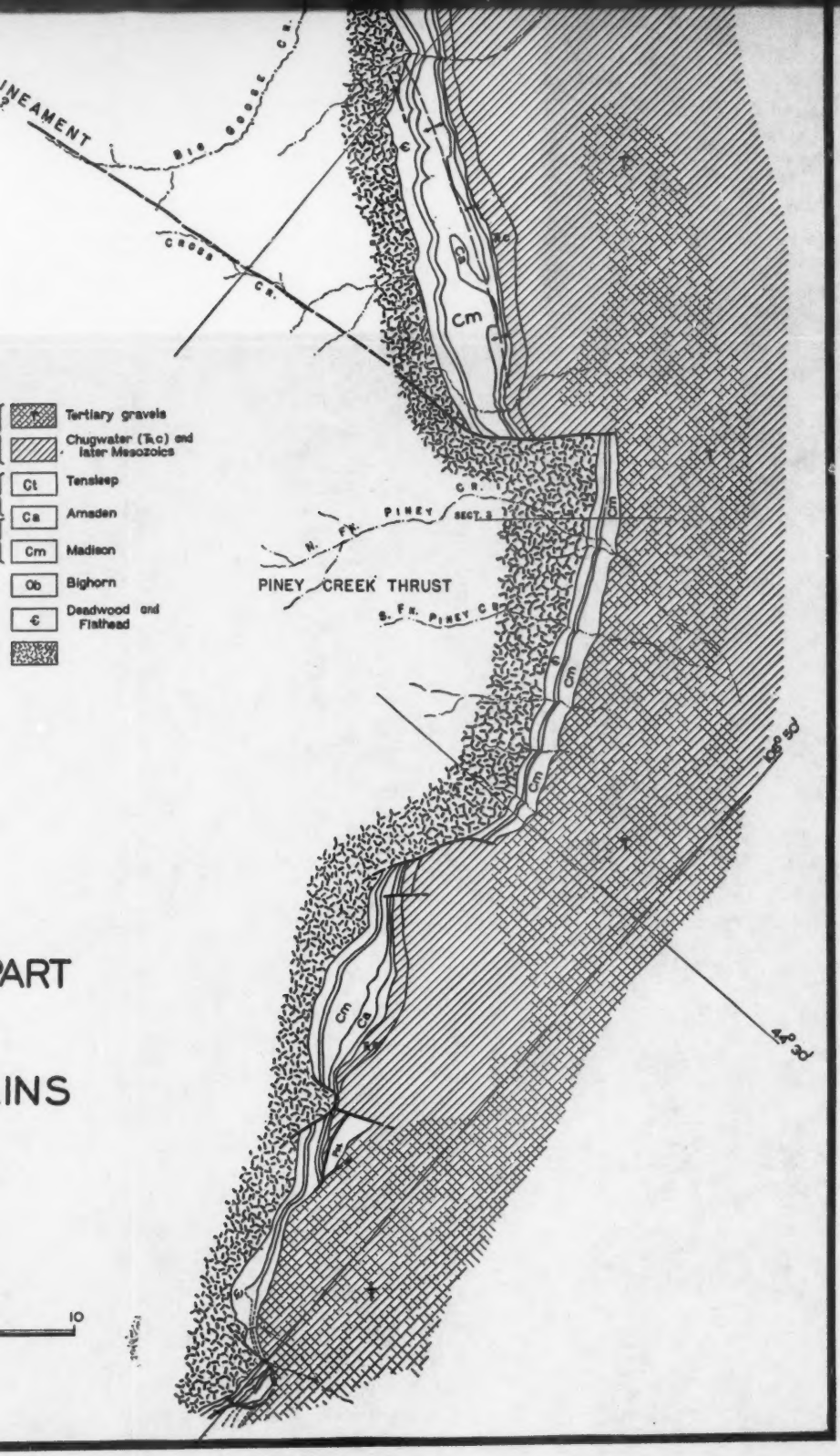

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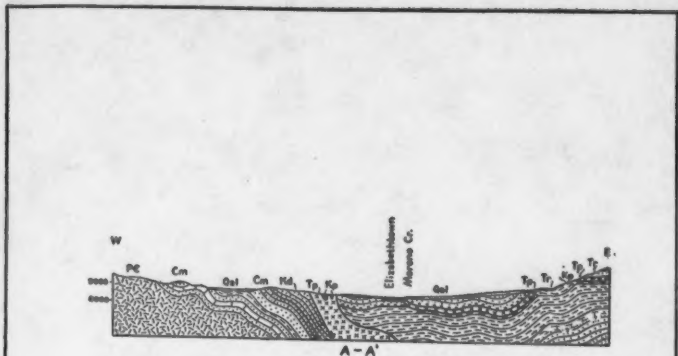
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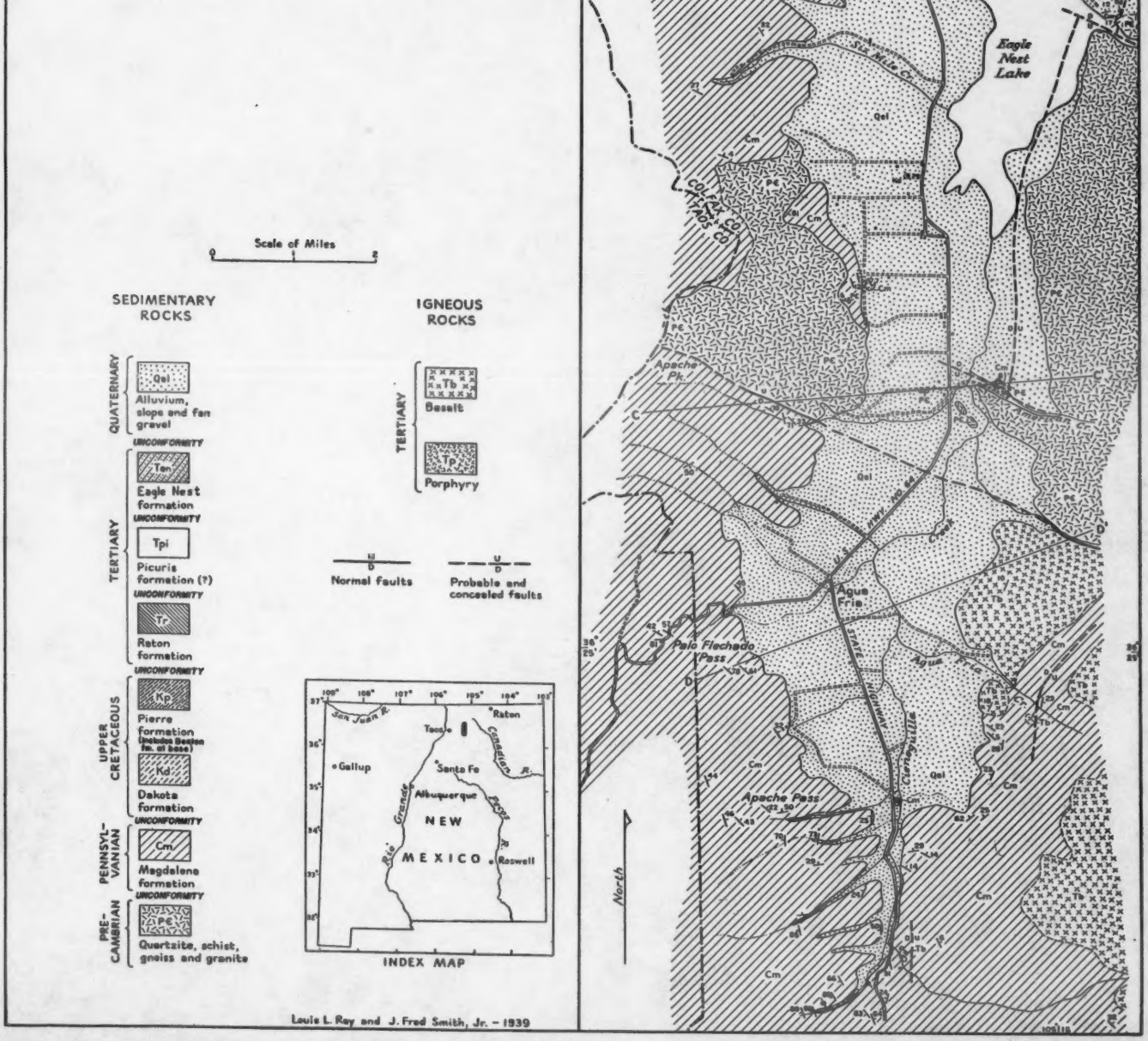

GEOLOGIC MAP AND STRUCTURE SECTIONS OF MORENO VALLEY

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GEOLOGY OF THE MORENO VALLEY, NEW MEXICO

BY LOUIS L. RAY AND J. FRED SMITH, JR.

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ABSTRACT

The Moreno Valley, located along the complex eastern boundary between the Sangre de Cristo Mountains and the Great Plains, is structurally a broad, northward-plunging syncline, disrupted by smaller folds and faults. This major synclinal structure is the result of the Laramide disturbance. Deformation, however, has continued possibly into the Quaternary.

Intrusions, probably contemporaneous with those of the Spanish Peaks, have deformed the strata in the northern portion of the Moreno Valley. Scully Mountain is a northwestward-plunging anticline produced by a porphyry intrusion.

A Mid-Tertiary surface of low relief is now represented by broad mountain flats at elevations near 10,000 feet. The Moreno Valley was initiated as a topographic feature about the close of the Mid-Tertiary. Its subsequent history is in many respects similar to that of the Rio Grande Depression.

During the Quaternary a broad valley stage and three lower pediments have been developed. The lowest pediments are cut on deformed valley fill, believed to be Pliocene and here named the Eagle Nest formation.

The early drainage of the valley was south and east into the Canadian River. Stream piracy by the Cimarron River diverted drainage to the east through the present Cimarron Canyon into the Canadian River. Drainage south of the point of capture has been reversed for about 14 miles, the present stream draining the southern portion of the Moreno Valley flowing northward to the Cimarron River.

INTRODUCTION

The Moreno Valley is the broad northern part of a topographic and structural depression in the mountainous portion of north-central New Mexico. It is nestled between the high Taos Range of the Sangre de Cristo Mountains on the west and the Cimarron Range on the east. The Taos Range culminates in Wheeler Peak (elev. 13,150+ feet). Peaks in the Cimarron Range rise to elevations of more than 12,000 feet. Relief in this region is great, for Baldy Peak in the Cimarron Range rises some 4300 feet above the village of Eagle Nest in a distance of about 7 miles, and Wheeler Peak rises more than 5000 feet above the village in a distance of 10 miles.

The depression, of which the Moreno Valley is the northern portion, strikes almost due north, parallel to the trend of the Sangre de Cristo Mountains, from the plains near Mora (Fig. 1) into the region of the high peaks near Costilla Pass (elev. 9700+ feet).

An unnamed divide (elev. 8851 feet) separates the Moreno Valley from the Valley of Coyote Creek, the southern part of the depression. The Moreno Valley has a maximum width of more than 5 miles and a length

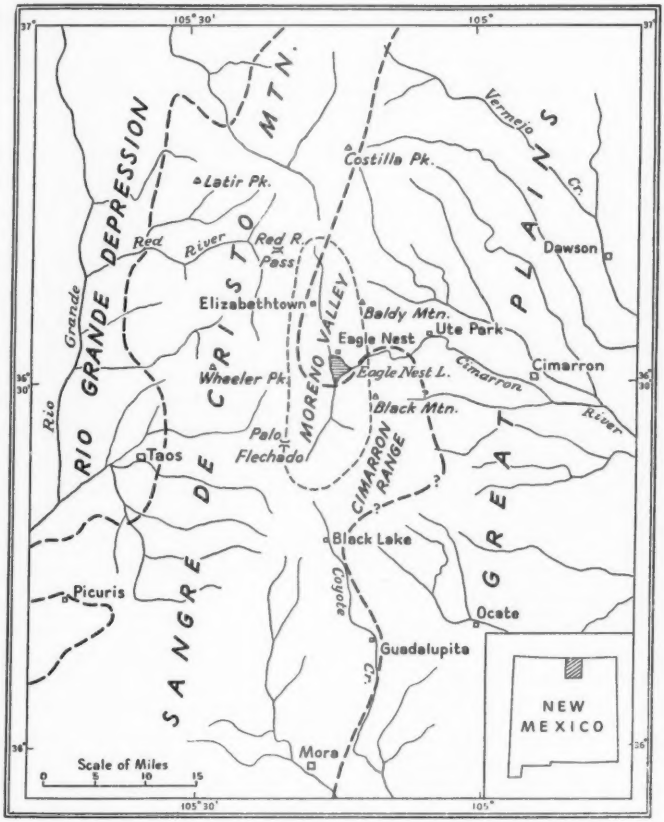


FIGURE 1.—*Index map*

Showing location of the Moreno Valley and its relationship to the regional physiographic provinces.

of about 25 miles, from Costilla Pass to the divide at the southern limit of the valley drainage. Midway between these divides, at an elevation of approximately 8000 feet, Moreno Creek, which drains the northern half of the valley, joins Cieneguilla Creek, which drains the southern half. The combined drainage of these two streams forms the Cimarron River,

which leaves the Moreno Valley through the picturesque Cimarron Canyon. The river follows a southeastward course to the Canadian River. At the head of the Cimarron Canyon, where the drainage leaves the Moreno Valley, a dam has been constructed which permits the flooding of a portion of the lowlands of the valley floor and the formation of Eagle Nest Lake (Fig. 1).

PREVIOUS GEOLOGIC INVESTIGATIONS IN THE MORENO VALLEY

The first known account of the Moreno Valley in geologic literature was by A. R. Conkling (1876), a geologist who accompanied the topographic party of the Wheeler Survey. He reported Cretaceous strata, from which two species of *Inoceramus* had been taken in the vicinity of Elizabethtown. He also observed Carboniferous strata at the Palo Flechado Pass (elev. 9107 feet), over which the present Taos-Raton highway passes (Fig. 1).

Stevenson's (1881) report for the Wheeler Survey contains the most complete account of the geology of this region, and the geologic maps accompanying that report are the only ones available for the Moreno Valley area. They served as the base for the geology of the Moreno Valley as shown on the small-scale geologic map of New Mexico, published under the supervision of N. H. Darton (1928), and here reproduced in simplified and modified form as Figure 2. When one considers the reconnaissance character of the work done by the Wheeler Survey, Stevenson's report stands as a monument to the excellence of field work done by the geologists of the early territorial surveys.

Except for scattered mining reports, many of which are unpublished (*see* Lindgren, Graton, and Gordon, 1910), no detailed geologic study has been made in the Moreno Valley since the days of the Wheeler Survey. Lee (1917; 1922; 1924) has mapped the Raton Coal Basin to the east, and Gruner (1920) has studied a portion of the Taos Range to the west, but neither has considered the geology of the Moreno Valley itself. Structure sections of the Moreno Valley, as shown in the report of Darton (1928, p. 273), rest primarily on the early reconnaissance studies by Stevenson.

Ellis (1931; 1935) reported a "glaciation" of the Moreno Valley. Ray (1940) has discussed this reported glaciation for which he could find no evidence.

SCOPE AND LIMITATIONS OF THE PRESENT INVESTIGATION

The writers spent several days in the Moreno Valley during the summer of 1938, when Ray was studying the glaciation of the Southern Rocky Mountains. They thought that knowledge of the structure and physi-

ography of this region, which was then little known geologically, was important for an understanding of the geologic history of the Sangre de Cristo Mountains.

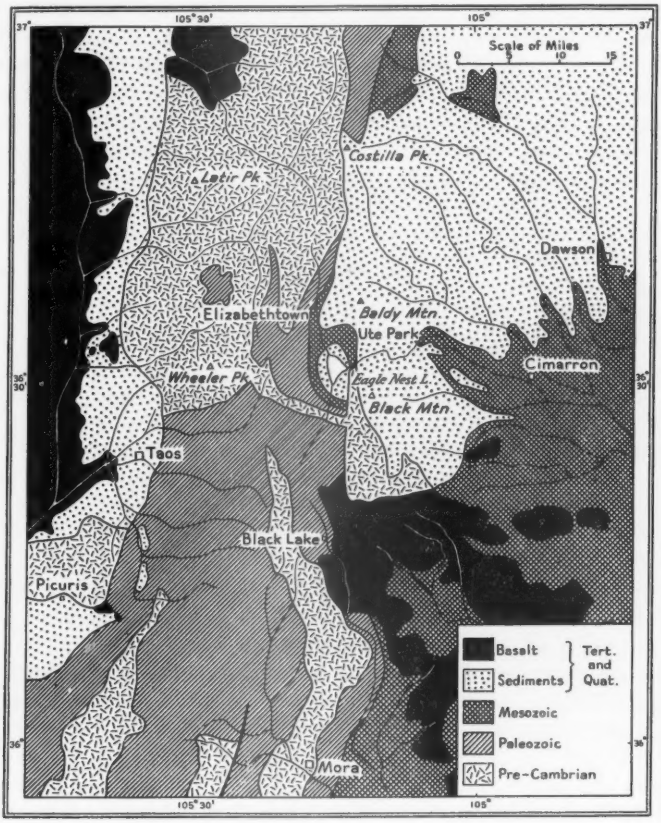


FIGURE 2.—*Generalized regional geologic map of Moreno Valley and adjacent regions*

Modified from Darton, Geologic map of New Mexico.

During the summer of 1939 10 weeks were spent mapping the geology and interpreting the physiographic history of this region. Adjacent areas which were of importance in the understanding and interpretation of the Moreno Valley were visited.

Unfortunately, continuous exposures of sedimentary rocks in the Moreno Valley are rare, and the stratigraphy of the Paleozoic and Mesozoic rocks is as yet little understood. Lack of exposures, irregularities in sedimentation, faulting, igneous intrusion, and the relative isolation of this region from areas of known stratigraphic sequence have made determinations of the displacements of faults and the amount of erosion along unconformities impossible to determine. Changes of sedimentary facies rendered correlation difficult. However, the writers believe that the general character of the structure is essentially as shown but that more detailed field work may indicate minor adjustments.

ACKNOWLEDGMENTS

The present study was financed by a grant from the Penrose Bequest of The Geological Society of America. The writers wish to acknowledge the hospitality of the many local people whose co-operation aided in the completion of this work. Especial thanks are due Mrs. O. F. Matkin of Laguna Vista Lodge, Mr. V. J. Van Lint of the Maxwell Land Grant Company, Mr. C. S. Heckt, mining engineer of the Aztec Mine, Baldy, New Mexico, and Mr. Charles Springer of the Springer Cattle Company. Mr. W. C. Patterson of the Ajax Mine furnished the writers with many unpublished mining reports. Mr. E. C. Cabot of Harvard University visited the writers and accompanied them to Mills Divide where a conference was held relative to the character of the gravel and sand deposits referred to the Picuris formation, originally described by Cabot (1938) from the western side of the Taos Range.

STRATIGRAPHY

GENERAL CONSIDERATIONS

The rocks of the Moreno Valley range in age from pre-Cambrian through Tertiary and into the Quaternary. Coarse red granite, granite gneiss, and white quartzite are the dominant pre-Cambrian rocks of the undifferentiated basal complex and the mountain core.

Sandstone, limestone, and conglomerate of the Magdalena formation (Pennsylvanian) rest unconformably on the pre-Cambrian. The Magdalena is in turn overlain unconformably by the coarse white Dakota sandstone, which, with the overlying soft black shales and thin limestone beds of the Benton and Pierre formations, is Upper Cretaceous.

Unconformably above the Pierre shale is the Eocene Raton sandstone and quartzite-pebble conglomerate. A relatively unconsolidated Tertiary formation, correlated with the Picuris, unconformably overlies the Raton and is tentatively assigned to the Late Miocene or possibly Early Pliocene.

Along the floor and eastern valley wall of the northern portion of the Moreno Valley are outcrops of an unconsolidated formation, composed of local angular to subangular debris, with interbedded sands, clays, and tuffs. This has been named the Eagle Nest formation and is believed to be Pliocene, with sedimentation possibly extending into the Quaternary. Unconsolidated Quaternary alluvium occurs along all the streams and capping Pleistocene pediments.

North of Eagle Nest Lake the sedimentary rocks of Eocene age and older have been cut by intrusions of monzonite porphyry and rhyolite. Basalt flows in the southern portion of the Moreno Valley region rest on an erosion surface of low relief high above the valley floor. Faulting has carried some of the basalt into the bottom of the valley. These lava flows are believed to be Mid-Tertiary.

PRE-CAMBRIAN

Pre-Cambrian rocks are best exposed along the eastern valley wall from Eagle Nest Lake dam to American Creek (Pl. 1). They consist primarily of a coarse red granite and a finer-grained granite gneiss, together with minor amounts of chlorite schist. North of Apache Peak, on the west side of the valley, a similar red granite crops out. The deeply weathered exposures in this region form pinnacled outcrops. Similar exposures occur in the vicinity of Comanche Creek and northward.

In the floor of the valley near the mouth of American Creek are several large outcrops of a coarsely crystalline pre-Cambrian quartzite, colorless to gray, or dark reddish-purple, which has been faulted and later folded. Curved fault surfaces on casual observation may be taken for roches moutonnées (Pl. 3, fig. 1). Similar quartzite has been described by Gruner (1920, p. 734-735), who has assigned it the name Pueblo quartzite. He believes it to be older than the intrusive granites.

No attempt was made by the writers to differentiate the various pre-Cambrian rocks in the field, and they have been mapped as a single unit (Pl. 1).

PALEOZOIC

Magdalena formation (Pennsylvanian).—The Magdalena formation crops out in the southern part of the Moreno Valley and along the western valley wall to a point about 1½ miles north of Comanche Creek (Pl. 1). It differs from the typical Magdalena in other portions of New Mexico in that sandstone predominates over limestone. It consists primarily of a conglomerate near the base, thin-bedded yellowish sandstone, shale, and limestone, with a red to deep maroon shale and sandstone near what is believed to be the top. The top has not been observed in the Moreno Valley region.

The conglomerate close to the base of the Magdalena is best exposed in the small creek valleys west of Scully Mountain (Pl. 1). It is moderately indurated with well-rounded phenoclasts of pre-Cambrian quartzite, granite, and granite gneiss, and a few cobbles of biotite schist. The cobbles, up to 18 inches in diameter, are surrounded by a coarse sandy matrix and are frequently fractured. In this vicinity the conglomerate is overlain by buff, greenish, and reddish limestone, shale, and sandstone with intercalated beds of gray limestone.

Typical sections of the lower portion of the Magdalena are exposed in road cuts at the Palo Flechado Pass (Pl. 1). Medium to coarse-grained yellowish sandstone beds of various thicknesses predominate in this vicinity. Some are ripple-marked and mud-cracked and contain abundant plant remains, among which *Calamites* is most common. Thin beds of arenaceous shale and limestone are interbedded with the sandstone. None of the strata could be traced far along the strike, and hence no serviceable horizon markers were found.

South of Agua Fria, massive coarse saccharoidal arkosic sandstone and conglomerate crop out. In part cross-bedded, they stand out prominently along the valley sides and floor (Pl. 3, fig. 2). Fossil plants are common in shaly lenses associated with these massive sandstones.

Red beds are found through much of the Magdalena section in this region and are particularly common in the upper portion. Along Agua Fria Creek (Pl. 1) is a thick series of light red to maroon sandstones and shales with occasional intercalated beds of yellowish sandy shale. Zones of greenish shale are irregularly interspersed. Similar strata are exposed beneath the Tertiary basalt flows at an elevation of about 10,000 feet along the southeastern part of the valley. Red beds crop out in the valleys west of Eagle Nest Lake, in the vicinity of Comanche Creek, and in the valley floor near the mouth of American Creek.

A distinctive limestone conglomerate, for convenience termed by the writers in the field the Bullington member of the Magdalena, lies immediately above the red beds. It is composed of subrounded prismatic fragments of limestone, some almost lithographic in character, embedded in a light-gray to buff matrix of compact limestone. The fragments have a maximum observed length of about 3 inches and a breadth of 1 inch. In places they are rare, and the pebbles are smaller, shapeless masses of light-buff limestone. Occasionally the conglomerate is slightly reddish. It is well exposed in a small outcrop north of the mouth of American Creek, near the Bullington Ranch House. It is common as float along parts of the west wall of the valley and crops out in the valley of Six Mile Creek, in the valleys south and west of Scully Mountain, and north of Comanche Creek. Field observations indicate that it lies stratigraph-

ically above all, or near the top of the "red beds" and below a sugary yellow sandstone which is presumably near the top of the Magdalena formation.

The total thickness of the Magdalena in the Moreno Valley has not been determined, for the complete section is not exposed. Lack of marker horizons makes it impossible to trace the strata from section to section. Darton (1928, p. 7) gives the average thickness as 600 to 1200+ feet. Mr. J. A. Young, Jr. (personal communication), who has made stratigraphic and paleontologic studies in the Taos Canyon and Black Lakes sections (Fig. 1), believes the minimum thickness to be more than 2000 feet. The writers believe that 2000 feet is a modest estimate, for the red beds along Agua Fria Creek in the upper portion of the formation have an estimated thickness of approximately 2000 feet. They probably comprise not much more than half the total thickness in this region.

The Magdalena rests unconformably on the pre-Cambrian rocks. The contact was observed only on the north side of Comanche Creek, where pre-Cambrian granite gneiss is overlain by a green conglomeratic sandstone containing feldspar and quartz pebbles up to three-quarters of an inch in diameter. Fractures in the pre-Cambrian rocks do not pass into the overlying sandstone and conglomerate. About 20 feet above the sandstone lies the coarse conglomerate previously described.

The Magdalena is overlain unconformably by the Dakota sandstone. Although the actual contact was not seen, the strata within a few feet above and below were observed in several places.

The Magdalena is Pennsylvanian in age, but it has not been dated more definitely. Paleontologic studies in the Taos Canyon and Black Lakes sections, adjacent to the Moreno Valley, have led Mr. J. A. Young, Jr. (personal communication), to the belief that this portion of the Magdalena may possibly be assigned to the Lower Pennsylvanian.

The massive sandstone south of Agua Fria was called Dakota by Stevenson (1881) and is so designated on the geologic map of New Mexico (Darton, 1928). Field relations suggest that it is a member of the Magdalena, probably occurring about the middle of the formation. Numerous well-preserved plant remains were found in an outcrop of this sandstone about 2 miles south of Agua Fria Creek. These were sent to Mr. C. W. Darrah of the Harvard Botanical Museum for identification. He reports the following fossil plants from this locality (Station 9-2, South of Agua Fria):

Neuropteris fimbriata, N. cf. *whitei (elrodi?)*
Alethopteris lonchitica
Asterophyllites radiatus
Odontopteris cf. *bertrandi*
Odontopteris genuina
Odontopteris obtusiloba
Cordaites
Calamites.

Darrah (personal communication) further states that this plant assemblage "belongs to the Later Pennsylvanian." If the tentatively identified *Neuropteris whitei* proved to be an *elrodi*, he would be "inclined to regard the beds somewhat older in the Pennsylvanian." It thus seems certain that the sandstones in the southern part of the Moreno Valley, formerly called Dakota, can be classed as Pennsylvanian rather than Cretaceous.

The possibility that the red beds overlying the massive sandstone assigned to the Pennsylvanian might be either Abo or Chupadera has been considered. However, studies by Mr. Young (personal communication) in the region of Black Lakes, immediately south of the area here considered, has substantiated the field conclusions of the writers that these strata are a part of the Magdalena formation. Young has found that south of the divide between Cieneguilla and Coyote Creeks the Lower Pennsylvanian guide fossil *Mesolobus mesolobus* occurs in "red beds" stratigraphically more than 500 feet above typical Magdalena sandstone, limestone, and shale. He believes that the red beds must be placed in the upper portion of the Magdalena formation. Almost vertical strata in the region just north of Black Lake (Fig. 1) show the transition between the characteristic fossiliferous Magdalena limestone, sandstone, and shale and the overlying red beds.

North of Idlewild (Pl. 1) sedimentary beds yielded fossil plants identified by Darrah as *Walchia piniformis*, which occurs in both the Pennsylvanian and Lower Permian. Plant fossils collected in the red beds of the Comanche Creek region belong to the same flora as that south of Agua Fria. Darrah (personal communication) has stated that he believes the whole flora falls within the Pennsylvanian, presumably in the upper portion, although the exact stratigraphic position within the Pennsylvanian is uncertain. Because of the lack of critical evidence, the writers hesitate to date the Magdalena more explicitly than Pennsylvanian.

MESOZOIC

Dakota formation (Upper Cretaceous).—The Dakota crops out in the region of Scully Mountain, north of Eagle Nest, and along the valley wall to the west of the village (Pl. 1). It is light gray, buff, or yellowish and unfossiliferous sandstone. The sand grains are angular to well rounded, and coarse to fine. In general the strata are massive and sugary, although some are quartzitic. Typically, thin films of quartz spread at random through the rock. They resist weathering and protrude from deeply weathered exposures, producing a cavernous or honeycombed appearance. The massive sandstone beds are prominent ridge formers.

At no place in the Moreno Valley could the thickness of the Dakota

sandstone be adequately determined, but the writers believe it to be considerably more than the 100+ feet stated by Lee (1917, p. 40). As shown on the map and sections (Pl. 1), it is probably nearer 500 to 700 feet in the vicinity of Scully Mountain.

The Dakota sandstone is the lower part of the Upper Cretaceous, resting unconformably on the Magdalena formation. At no place in the Moreno Valley was the upper contact of the Dakota observed, but it is apparently conformably overlain by the Benton formation.

Benton formation (Upper Cretaceous).—Black shale and thin beds of limestone compose the Benton formation in the northern portion of the Moreno Valley, near Elizabethtown. The writers first believed that all the beds above the Dakota and below the Raton belonged to the Pierre. However, near the base of these strata occur numerous *Inoceramus labiatus,* a characteristic Benton fossil.

There is no apparent lithologic break between these lower beds and the overlying strata, which have been mapped as Pierre. For this reason, and because of the lack of outcrops, the Benton has been included in the Pierre on the geologic map (Pl. 1). It was not possible, because of lack of exposures, to determine whether there were any strata of the Niobrara formation between the Benton and the Pierre.

Pierre formation (Upper Cretaceous).—The Pierre, which is exposed in the Moreno Valley only to the north of Eagle Nest Lake, is lithologically similar to the Pierre of the plains to the east. It is a compact black shale, containing occasional thin layers of limestone, calcareous shale, arenaceous shale, and septarian nodules. Cone-in-cone structure was found in some of the black shale beds. The calcareous beds occur near the base.

In the vicinity of Elizabethtown and Iron Mountain intrusive dikes and sills have metamorphosed and indurated the shale, and in places replaced it by siliceous skarn minerals. Some of the metamorphosed shale resembles a light-gray, fine-grained quartzite.

Because of intrusion, faulting, and the unconformity between the Pierre shale and the overlying formations the complete thickness of the Pierre formation in the Moreno Valley was not obtained. Lee (1917, p. 40) gives 3000+ feet as the thickness of the Benton, Niobrara, and the Pierre formations in the Raton Coal Basin. Darton (1928, p. 7) does not give the thickness of the Pierre in northeastern New Mexico, but Knopf (1936, p. 1731) indicates that there is more than 1000 feet of Pierre shale in the Spanish Peaks region, northeast of the Moreno Valley. At Mills Divide, called Ponil Pass by Stevenson (1881), 500 feet of the Pierre shale is exposed. It does not include, however, the indeter-

minate amount of igneous intrusive rocks which occur in the section. Neither the top nor the base of the formation can be observed at this locality.

The Pierre shale is believed to rest on the lithologically similar Benton formation and is in turn overlain unconformably by the Raton. The Benton has been separated from the Pierre on paleontological evidence alone. Later Cretaceous formations were presumably deposited in this region but were eroded before the deposition of the Raton. Lee (1917, p. 67) stated that in the Baldy region, just east of the Moreno Valley, the basal conglomerate of the Raton is variable and is obviously unconformable on the Pierre formation. In the Pierre at Mills Divide, the writers found poorly preserved fish scales, fragmentary shells of *Inoceramus* sp., and a single specimen of an ammonite, tentatively identified as *Placenticeras* sp.

TERTIARY

Raton formation (Eocene).—The Raton crops out in the Baldy Mountain region east of Elizabethtown, in the vicinity of Elizabethtown, and at Mills Divide (Pl. 1). None of these exposures is extensive, nor do any show great thickness. The base of the formation, cropping out immediately to the west and north of Elizabethtown, is a coarse 50-foot quartzite conglomerate resting on the Pierre shale (Pl. 3, fig. 3). The white quartzite cobbles, whose average diameter is about 3 inches and whose maximum diameter is about 8 inches, are similar to and were probably derived from the pre-Cambrian quartzite to the north. The matrix is coarse white to yellowish sand.

Above the conglomerate an indeterminate thickness of coarse yellowish and white friable sandstone contains occasional thin carbonaceous layers. The sandstones of the Raton are poorly exposed. Conglomeratic layers resist weathering and tend to form prominent exposures.

Because of the few exposures and limited vertical range, the thickness of the Raton formation in the Moreno Valley was not obtained. Lee (1917, p. 40) has reported it to be 1800± feet in the Raton Coal Basin.

The Raton formation lies unconformably on the Pierre shale, the intervening Trinidad and Vermejo formations having been eroded prior to the deposition of the Raton. The contact between the Raton and the Pierre is exposed for only a few feet laterally so that no information could be obtained other than that the quartzite pebble conglomerate rests on black to yellowish, arenaceous shale of the Pierre formation.

The upper contact of the Raton formation is not exposed in the Moreno Valley. At Mills Divide about 25 feet of a younger Tertiary formation occurs stratigraphically not more than 100 to 300 feet above

the basal Raton conglomerate. This indicates a long period of erosion between the deposition of the Raton and that of the overlying Tertiary formation.

Since no fossils were found in the Raton formation in this area, its age cannot be given more definitely than that it is post-Pierre and is cut by intrusive igneous rocks. Lee (1917, p. 60-61) has called this formation Eocene in the Raton Coal Basin: "The paleontology is in perfect accord with the structure in indicating the early Tertiary age of the Raton formation." However, later paleobotanical evidence, based on long-range correlation, has led to the belief that the Raton formation is pre-Eocene. Knopf (1936, p. 1731) has called it "probably of late Cretaceous age" in the Spanish Peaks region. Recently, Chaney (1940, Fig. 1) has classed the Raton formation as Eocene. Because of the stratigraphic and structural relationships the writers have supported the Eocene age.

Picuris (?) formation (Late Miocene?).—The strata here referred to the Picuris formation were originally described by Cabot (1938) in the Taos region to the west. They are exposed at Mills Divide (Pl. 1; Pl. 4, fig. 1) at an elevation of more than 9750 feet. Remnants of a more or less unconsolidated red sandstone, red shale, greenish shale, and coarse sandy and conglomeratic layers crop out in a narrow gully formed by the overflow from an old flume. The entire outcrop and overlying soil cover are stained a deep maroon. Flat angular fragments of the local intrusive porphyry were found in the conglomeratic layers.

The most striking property of the Picuris (?) in this area is the fractured and deformed cobbles of pre-Cambrian quartzite scattered through coarse reddish sand. Though deformed, the broken parts of the cobbles are in place, and abrasions show where adjacent cobbles were mashed against one another. They are smooth and rounded, indicating that they were well worn during transportation.

The Picuris (?) is post-intrusive. Except for physiographic reasons, no more definite age can be assigned to the formation, which has thus far yielded no fossils. Physiographic data engender the belief that it is post-mid-Tertiary, probably latest Miocene or very early Pliocene. In the Rio Grande Depression, Cabot (1938, p. 91) has suggested that the Picuris formation is pre-Santa Fe in age.

The beds at Mills Divide have been correlated with the Picuris largely on the basis of lithology. The red sands and fractured cobbles are typical of the lower portion of the Picuris to the west. The section at Mills Divide is believed to be near the base. No evidence was seen which would in any way suggest fault relations between the Picuris and the underlying Raton.

It is believed that at the time the Picuris (?) was deposited the higher mountain summits were Mid-Tertiary monadnocks. In the vicinity of the Moreno Valley the deposits represent the deposition by swiftly flowing streams, either confined in valleys or spread out on the erosion surface at the base of the monadnocks. Lack of exposures prevents a clearer picture concerning the origin of this formation and its regional relationships. Later deformation, probably accomplished at the time the Moreno Valley was outlined by faulting, fractured the cobbles.

Eagle Nest formation (Pliocene?).—Along U. S. Highway 64, north of Eagle Nest Lake are outcrops of an unconsolidated, deformed, and faulted Tertiary formation here called the Eagle Nest. Reddish clay and white tuffs are interbedded with coarse white to buff sand and gravel. Cobbles up to 8 inches in diameter are not uncommon. One sandstone boulder measured 18 inches across. In general the cobbles are smooth but subangular and are composed primarily of the local porphyry and rhyolite, sandstone, and fragments of metamorphosed Pierre sale. A few small channel deposits of sand were seen in the thicker beds of clay.

Along the north shore of Eagle Nest Lake, west of the dam, a small ripple-marked outcrop of a fine-grained white sandstone was seen. It is believed to be a part of the Eagle Nest formation. East of Elizabethtown, deep gullies, cut during placer mining operations, expose sections of unconsolidated sand and gravel of local origin which are believed to be correlative with the outcrops along the highway at Eagle Nest Lake (Pl. 4, fig. 2).

Numerous small faults occur in the Eagle Nest formation at the type locality, where the beds dip approximately 21 degrees toward the center of the valley. Attempts to obtain the thickness of this formation were unsuccessful since neither the top nor the base has been seen.

No fossils were found in the Eagle Nest formation, but, because of its position near and along the floor of the valley, and because of its lithology, it is believed to be younger than the Picuris (?). These strata are tentatively assigned to the Pliocene and perhaps early Pleistocene. This age determination is based primarily on physiographic evidence, to be discussed later.

The Eagle Nest formation represents a valley filling, or basin filling of Late Tertiary and possibly later age. It was deposited within the Moreno Valley after the formation of the valley depression by faulting. The formation probably represents great coalescing stream fans, perhaps similar to the present alluvial fan of Willow Creek (Pl. 2), which choked the Moreno Valley with locally derived debris to a height determined

by the gradient of the integrated drainage of the valley. The total thickness of the formation must therefore have been approximately 1000+ feet (Pl. 2).

The origin of the Eagle Nest formation is believed to have been similar to that postulated for the Santa Fe formation in the Rio Grande Depression (Bryan, 1938, p. 205; Denny, 1940). They have shown the Santa Fe to be a basin filling of alluvium, deposited mainly as coalescing fans. Tuffaceous material within the formation indicates volcanism during the time of deposition.

The writers believe that much of the Moreno Valley is underlain by the Eagle Nest formation, but covered by Quaternary pediment gravel and alluvium. Therefore, only exposures have been mapped, and no attempt has been made to show the areal extent of the formation (Pl. 1).

The name Eagle Nest is applied to this unconsolidated Late Tertiary formation in the Moreno Valley because of the proximity of the best exposures to Eagle Nest Lake, along U. S. Highway 64.

QUATERNARY DEPOSITS

An unknown thickness of recent alluvium covers large portions of the floor of the Moreno Valley. It is composed of coarse sand, gravel, and fine black silt. Wide grassy meadows developed on the black soils of the flood plains are termed *vegas*, from the Spanish meaning "fertile plain." Many of the smaller tributary creeks, especially south of Agua Fria, occupy valleys with broad floors, or *vegas*, much wider than can be justified by the present size of the streams (Pl. 5, fig. 1). Pediments rising steplike throughout the valley are capped by coarse gravel, poorly consolidated by caliche near the surface.

Much of the Pleistocene and recent gravel is primarily composed of reworked debris from the Eagle Nest formation, from which it is distinguished only with difficulty. Thus, water wells drilled in unconsolidated debris to depths of over 100 feet in the floor of the valley (Mr. W. A. Thomas, personal communication) are of little value in the determination of thickness of either the Quaternary or the Eagle Nest formations.

IGNEOUS ROCKS AND MINERALIZATION OF THE MORENO VALLEY

All the formations in the Moreno Valley, up to and including the Raton formation, have been cut by intrusive igneous rocks. Dikes and sills are especially common in the region north of Eagle Nest Lake, where the sedimentary rocks have been deformed and metamorphosed by the intrusives (Pl. 1). Mineralization has occurred in association with many of the intrusive bodies (Lindgren, Graton, and Gordon, 1910, p. 82-108). Gold and copper have been mined in the Moreno Valley and adjacent

regions. In 1939 a little gold was being mined, and a small tungsten prospect had been opened on the southeastern side of Iron Mountain. A few miles to the west, in the Red River Valley, molybdenum, associated with Tertiary intrusives, is being mined (Larsen and Ross, 1920; Vanderwilt, 1933).

The intrusive rocks of the Moreno Valley may be classified in two major groups: rhyolite and quartz monzonite porphyry. The rhyolite is usually white to buff with a very fine texture. Quartz phenocrysts are small, and mafic minerals rare. The rhyolite forms small irregularly shaped bodies cutting Mesozoic sediments along the eastern valley wall near Eagle Nest Lake dam and north of the highway (Pl. 1). Along the north wall of Comanche Creek Canyon, about 1½ miles above its mouth, small irregular bodies of rhyolite cut pre-Cambrian granite and granite gneiss. To the south, near the Klondike Mine at Idlewild, numerous small rhyolite bodies occur. These outcrops are too small to be shown on the geologic map. Many prospects have been opened near these outcrops, but no mineralization of sufficient value has been discovered to warrant mining.

The most important intrusive rocks are the quartz monzonite porphyries which are common throughout the Elizabethtown district. The texture of the porphyry may vary, but considered as a whole there is a singular uniformity of appearance in the many dikes and sills of this rock type. The larger bodies tend to be in general more porphyritic than the smaller, with phenocrysts of feldspar and quartz occasionally attaining diameters of more than a centimeter. At the head of Willow Creek, east of Iron Mountain, the porphyry has a finer and more uniform texture. It frequently contains much magnetite. In the Cimarron Canyon there is exposed, just east of the Moreno Valley, a large sill-like body of the quartz monzonite porphyry, about 400 feet in thickness.

Where the porphyry is in contact with the Pierre shale, considerable contact metamorphism has occurred, frequently with the introduction of gold and the production of skarn minerals, especially epidote. In Willow Creek valley east of Iron Mountain where the best examples of contact metamorphism are exposed the Pierre shale near the contact has been highly metamorphosed with the production of skarn minerals. There is a pronounced baking of the shales, which are frequently so silicified that on casual observation the rock might be mistaken for a quartzite.

The Aztec Mine on the eastern slope of Baldy Mountain is a good example of a characteristic mineral deposit formed where mineralizing solutions spread from the intrusive mass and penetrated the country rock,

occasionally depositing rich concentrations of gold (Lee, 1916; Chase, 1922).

Near the mouth of Agua Fria Canyon and at Mills Divide (Pl. 1) small sills are exposed. The first, intruded into the Magdalena formation, is so highly weathered that no attempt has been made to name the rock. At Mills Divide, a dacite porphyry sill, slightly more than a foot thick, intrudes the Pierre shale. Contact metamorphism at both localities is almost negligible.

The period of intrusion cannot be closely dated, but it is post-Raton and pre-Picuris (?), for cobbles of the intrusive rocks are found in the Picuris (?) formation. Emplacement was prior to the Mid-Tertiary faulting which outlined the Moreno Valley. It is possible that the igneous activity in the Moreno Valley region is closely related to that in the Spanish Peaks region, some 60 miles to the north and east, recently described in detail by Knopf (1936).

South of American Creek, within the valley and above the eastern valley wall, are extensive basaltic lava flows. In general the basalt is homogeneous. However, occasional porous zones and frequent amygdules of calcite are present. No attempt was made to differentiate between the various lava flows. It should be noted, however, that several of the flows contain phenocrysts of quartz and may be termed quartz basalt.

The basalt flows are older than the major faulting, which outlined the Moreno Valley, but younger than some of the minor faulting. Field relations indicate that they are somewhat younger than the intrusive rocks and are related to the earliest lava flows of the Ocaté region, with which they are continuous (Fig. 2).

STRUCTURE

GENERAL STATEMENT

The major structural feature of the Moreno Valley is a broad northward-plunging syncline, broken by transverse faults and faults parallel to the axis of the syncline. Numerous small, sharp folds are superimposed on the major syncline in the southern part of the valley and in the vicinity of Elizabethtown. In the latter region they appear to be in many instances the result of deformation by igneous intrusion. The major axial faults are on the east side of the valley and trend north-south, parallel to the valley and to the main structural features of the Sangre de Cristo Mountains. A smaller fault with a similar trend occurs on the west side of the valley. Transverse faults cut the Moreno Valley syncline south of Eagle Nest Lake and through the canyon in which the dam is located at the source of the Cimarron River (Pl. 1).

FOLDS

The Moreno Valley is underlain by a large northward-plunging syncline, whose axial trend is approximately along the center, or slightly east of the center of the valley. Dips on the west limb of this syncline range from 20° to 70° to the east. Because of the plunge of the syncline, which is about 8° to 14° to the north, the beds strike to the northwest on the west limb and to the northeast on the east limb. This synclinal structure appears to extend throughout the entire length of the valley, so that successively younger strata crop out from south to north. The regular stratigraphic succession, however, has been broken by transverse faulting.

Numerous small folds indicating complex deformation in the southern portion of the valley represent minor wrinkles on the limbs of the large syncline. As outcrops are scarce and there are no good horizon markers, not all these folds can be mapped in detail. No attempt has been made to show the complexity of this folding on the geological map (Pl. 1), for its significance in a study of this nature is limited. Many beds dip steeply, and in places small folds plunge as much as 50° north. Across the divide at the southern end of the Moreno Valley deformation has been even more intense. Between this divide and Black Lakes (Fig. 1) many beds are vertical and in places may be slightly overturned.

Scully Mountain (Pl. 1) is structurally a northwestward-plunging anticline with a core of intrusive quartz monzonite porphyry. The anticline appears to have been the result of intrusion of the porphyry which has warped and uplifted the Dakota sandstone. Dips on the flank of this structure average about 30° to the east on the east side, approximately 15° to the northwest on the northwest side, and about 55° to the southwest on that side. The structure is probably domal, but, as the strata on the southeast have been removed by erosion, the attitude of the beds cannot be determined.

Along the north side of Comanche Creek are several small, northward-plunging anticlines and synclines. Iron Mountain is anticlinal (Pl. 1, Sec. B-B′). Numerous small folds in the vicinity are difficult to delineate because of lack of outcrops. Many of these smaller folds are the result of local doming by the numerous intrusives. Some, however, may be due to early Tertiary deformation. Lee (1922) has shown that in the Raton Coal Basin the Eocene Raton formation has been warped into a broad syncline. It is not known how much of the deformation in the vicinity of Elizabethtown may have been due to this period of deformation and how much has been produced by the igneous intrusions.

FAULTS

Most of the faults of the Moreno Valley trend north-south, parallel to the axis of the valley. Those mapped are mainly along the sides of

the valley. Alluvium in the center and along portions of the valley walls may conceal numerous faults.

A large fault or fault zone, trending slightly east of north, extends along the east side of the Moreno Valley from Eagle Nest Lake southward (Pl. 1). The west side is downthrown, and the Magdalena formation is brought in contact with the pre-Cambrian to the east. At no place along the southern part of the lake region is this fault actually exposed, but in a small outcrop 1¼ miles south of the lake Magdalena beds strike directly into the pre-Cambrian about 100 feet to the east, indicating that the east limb of the large syncline has been broken by the fault. Several springs occur along the fault zone between the south end of Eagle Nest Lake and the mouth of American Creek. Just north of the lake, Cretaceous beds appear to be in contact with the pre-Cambrian along this same fault, which has been slightly offset by a transverse fault (Pl. 1).

Along the north shore of the lake, a few hundred feet above the dam, is a zone of faulting and brecciation about 300 feet long. This fault zone continues to the northeast, although the fault plane is not visible, for pre-Cambrian granite and granite gneiss lie to the east and the Magdalena sandstone and conglomerate and perhaps younger strata lie to the west. Near the faults on the lake shore is a small outcrop of a fine, white, ripple-marked Tertiary sandstone, believed to belong to the Eagle Nest formation. Its relation to the faults could not be determined.

Since it was impossible to obtain the thickness of the Magdalena formation the throw of this major fault is not known. South of Eagle Nest Lake the Bullington conglomerate member near the top of the Magdalena formation is faulted against pre-Cambrian quartzite. As the Magdalena formation seems to be several thousand feet thick in this region, the throw of the fault is at least as great as the local thickness of the Magdalena formation. Besides the Magdalena, north of the lake are small outcrops of sandstone which are probably Dakota. As these strata contain no fossils and occur as small fault blocks, their age and the throw of the fault have not been determined.

The course of the fault north of the lake was not followed into the mountains. The eastern valley wall in the vicinity of Elizabethtown is probably the result of minor folding (Pl. 1, Sec. A-A′). Graton (Lindgren, Graton, and Gordon, 1910, p. 94, Fig. 4) in his reconnaissance study of this region has shown a diagrammatic section through the Elizabethtown district, in which he indicated probable faults. Since no faults of this character were observed by the writers, the structure has been interpreted

as primarily folded. Deformation by folding accords with the local geology. Sandstone of the Raton formation caps Baldy Mountain at an elevation of more than 12,000 feet, and the basal conglomerate occurs at Elizabethtown, 3 miles to the west and more than 3500 feet lower. The thickness of the Raton formation is given by Lee (1917, p. 40, 57) as 1800± feet, with a maximum thickness of about 2000 feet in the Raton Mesa. An indeterminate thickness of porphyry within the formation as well as the observed dips suggest deformation by folding rather than faulting. This does not mean, however, that no faulting has occurred, merely that faulting is locally subordinate to folding.

Several transverse faults cut the main fault on the eastern side of the Moreno Valley. A transverse fault trending northwest passes through Eagle Nest Lake (Pl. 1). No fault plane was observed here and the fault has been shown because it explains the offset of the north-south fault along the east side of the lake, accounting for the fact that exposures on the south side of the narrow channel at this point are pre-Cambrian, whereas on the north side of the channel there are outcrops of Magdalena and younger formations. This postulated fault zone also explains why the Cimarron River tapped the drainage of the Moreno Valley through a channel of pre-Cambrian rocks instead of a channel cut in less competent rocks such as the Mesozoic and unconsolidated Tertiary sediments a few hundred feet to the north. The river has followed a zone of structural weakness.

South of Eagle Nest Lake, along American Creek (Pl. 1) the main fault is broken by a cross fault which trends about N. 60° W. The north side is downthrown, probably several thousand feet, bringing the upper part of the Magdalena formation on the north in contact with the pre-Cambrian quartzite on the south. Another transverse fault 1¼ miles to the south also trends about N. 60° W., and its south side is downthrown. This fault brings Magdalena sandstone on the downthrown side in contact with pre-Cambrian granite on the north. On the east side of the Moreno Valley, Tertiary basalt is faulted against pre-Cambrian granite. Again the throw cannot be determined as the exact stratigraphic horizons within the Magdalena are not known. Along the east side of the valley the basalt rests directly on the Magdalena formation, so the fact that the Tertiary and pre-Cambrian rocks are in contact does not add appreciably to the amount of throw of this fault. At no place is the actual fault contact exposed, but at several localities on the west side of the valley the rocks crop out on both sides within a few feet of the fault.

Outcrops of pre-Cambrian quartzite near the mouth of American Creek show numerous curved, smooth, slickensided surfaces (Pl. 3, fig. 1).

Whether these were formed at the same time as the larger faults is unknown, for no contact of these quartzite outcrops with later formations is exposed. The faulting could be as old as pre-Cambrian.

West of Eagle Nest Lake a fault trends slightly west of north, the west side downthrown, so that Dakota sandstone is in contact with the Magdalena formation (Pl. 1). This fault is terminated on the north by an east-west cross fault whose south side is downthrown, so that the Dakota is in contact with the Magdalena to the north. The throw along both these faults is not more than a few hundred feet, as the normal contact between the Magdalena and Dakota occurs only a mile west of the north-south fault.

In the northern portion of the valley a north-south fault follows North Moreno Creek (Pl. 1), and the Dakota sandstone on the east is downthrown against pre-Cambrian rocks on the west, indicating a throw of several thousand feet.

In the southern portion of the valley two small faults are indicated to explain the position of the Tertiary basalt. Along Agua Fria Creek the basalt within the valley dips about 10 degrees toward the valley center, while on the highland above the valley it is almost horizontal. A wide valley cut through the basalt into the Magdalena is probably developed along the zone of faulting.

Numerous small faults have been observed in the pre-Cambrian formations of the Moreno Valley, as along Comanche Creek. It was not possible to trace them far, and as they are of minor significance in the present study they have not been shown on the geologic map.

The faults of the Moreno Valley are probably normal, although none of the fault surfaces are exposed. The nature of the faulting in near-by regions and the character of the outcrop pattern is the main evidence used in deducing the character of the faulting.

REGIONAL STRUCTURAL RELATIONS

The fault pattern of the Moreno Valley is similar to that along the east side of the Rio Grande Depression, west of the Sangre de Cristo Mountains. Cabot (1938) shows a discontinuous fault boundary between the mountains and the depression similar to the boundary between the Moreno Valley and the Cimarron Range. The same pattern occurs along the boundary between the Culebra Range of the Sangre de Cristo Mountains and the San Luis Valley (Upson, 1939).

Faults approximately parallel to those along the east side of the valley probably continue to the south at least as far as Mora (Fig. 2). The area of pre-Cambrian rocks east and southeast of Eagle Nest Lake (Fig. 2; Pl. 1) is believed by the writers to be essentially a fault block in

a region which is otherwise a part of the Great Plains physiographic province. The structural relations along the north and south ends of this block are not known.

The structural history of the Moreno Valley is essentially similar to and contemporaneous with that of the surrounding regions.

TIME OF FOLDING AND FAULTING

Excepting the recent alluvium, all the rocks in the Moreno Valley have been deformed several times.

Pre-Magdalena diastrophism (presumably pre-Cambrian) is indicated by the metamorphosed rocks and by the faulting and upturning of beds of the pre-Cambrian quartzite near the mouth of American Creek. The near-by Magdalena strata have suffered a minimum of folding.

No angular unconformity has been observed between the Paleozoic and Mesozoic strata, which were probably folded by the Laramide disturbance when the synclinal structure of the Moreno Valley was produced. This deformation may have been accompanied by some faulting, but the writers believe that most of the faulting in this region is later.

The major faulting which outlined the Moreno Valley was initiated after (1) the igneous intrusions, (2) the formation of a Mid-Tertiary surface of low relief, and (3) the deposition of the Picuris (?) formation (*see* section on geologic history) and is presumably of late Miocene or early Pliocene age. Preceding it a period of folding produced the syncline of the Raton Coal Basin (Lee, 1922, Fig. 5). This deformation is post-Raton and pre-Picuris (?). No evidence of structural deformation which can be assigned to this period has been found in the Moreno Valley.

Following the deposition of the Eagle Nest formation in the fault basin there has been a continuation of movement. Along Highway 64, north of Eagle Nest Lake, the formation dips as much as 21° to the west. At first this was thought to be entirely initial, but after examination of the post-Eagle Nest faulting it was concluded to be an initial dip, exaggerated by deformation. Numerous small faults are exposed throughout this formation along Highway 64, at the type locality. They constitute the youngest faulting seen in the valley and are Pliocene or Quaternary. No recent fault scarps were observed within the valley, but Upson (1939, p. 731) has described small ones cutting alluvium along the western front of the Sangre de Cristo Mountains near the Colorado-New Mexico State boundary. It is believed that faulting has continued intermittently in the Moreno Valley from the period of the Laramide deformation to recent times.

PHYSIOGRAPHY OF THE MORENO VALLEY

RELATIONSHIP TO MAJOR PHYSIOGRAPHIC PROVINCES

Physiographic boundaries previously drawn in northeastern New Mexico (Lee, 1922; Fenneman, 1931; Bryan, 1938) have been too generalized to show adequately the boundary between the Southern Rocky Mountains and the Great Plains. To the north, in Colorado, the separation is made on the simple basis of structure and topography. In northern New Mexico this simple separation does not suffice, for here the gently dipping Tertiary rocks of the plains reach elevations of 12,000 feet west of the Raton Coal Basin. The problem, then, is where the mountains and plains should be separated, for topographically the division may be made at one place, structurally at another. If one is to draw the boundary between the mountains and the plains on the basis of topography alone, the Moreno Valley, which lies wholly above 8000 feet, must be considered a part of the mountain province. However, structurally the northern portion of the Moreno Valley lies within what is generally conceded to be the Great Plains Province, for here the hogback-forming strata crop out along the western valley wall, and gently eastward-dipping Mesozoic and Tertiary strata form the eastern valley wall (Pl. 1). To the south these formations do not occur in the Moreno Valley, so that the valley lies along a structurally complex and irregular boundary.

The line of separation between the Southern Rocky Mountains and the Great Plains is further complicated by the fact that the mountains east and west of the Moreno Valley, the Cimarron and Taos ranges of the Sangre de Cristo Mountains, are physiographically similar. Relief is great, stream gradients steep, and peaks rise to elevations of 12,000 feet or more. Structurally, however, the mountain ranges are dissimilar. The Taos Range is a part of the complexly folded and faulted geanticline of the Southern Rocky Mountains. The Cimarron Range is a dissected, eastward-tilted fault block portion of this same geanticline.

Near Mills Divide, on the north flank of Baldy Mountain (Pl. 1), there is no indication of a structural break between the gently upwarped sediments of the Great Plains and those of the Moreno Valley. Elevation is the only reason for considering this a part of the mountain province. Lee (1922) found some difficulty in distinguishing between the mountains and plains and in the Raton Coal Basin he referred the summits to the Great Plains Province. He called attention to the fact that these "mountains", while structurally a part of the Great Plains, are composed of gently dipping and dissected Tertiary and Mesozoic sediments, with peaks rising more than 10,000 feet. This is, according to

Lee (1922, p. 1), the highest and most intensely dissected portion of the Great Plains.

After a careful consideration of the merits of both the topographic and structural separation of the mountains and plains, the writers have concluded that the Cimarron Range should be considered a fault block extension of the Southern Rocky Mountains, surrounded on all but the west side by the Great Plains. Boundaries of this block are only tentative (Fig. 1). This interpretation results in the classification of the Moreno Valley as a border depression along the eastern margin of the Southern Rocky Mountains, which in this region is complicated by an eastward extension of the mountains as the Cimarron fault block. North of the Cimarron Range the Great Plains extend to the upturned hogback formations along the floor and western wall of the Moreno Valley. South of the Cimarron Range the boundary cannot be defined because of the cover of Tertiary lavas (Fig. 2). It lies, however, east of the present valley. Farther south, near Guadalupita and Mora, the boundary line is drawn at the contact of the Paleozoic and Mesozoic sediments (Fig. 1, 2).

PHYSIOGRAPHIC DEVELOPMENT

General statement.—The Moreno Valley owes its peculiar position and topographic form to the complex interrelationship of structure and physiographic processes. The primary position and general outline of the valley have been determined by the underlying bedrock, and stream development has given rise to the secondary topographic forms within the valley.

At the close of Mid-Tertiary time the Moreno Valley area was part of a widespread erosion surface. To the east and west the higher peaks of the present Cimarron and Taos ranges rose as monadnocks (Pl. 6). Post-Miocene normal faulting, previously described, disrupted the continuity of this erosion surface at the site of the Moreno Valley, with the production of a warped fault trough or basin. Erosion of the highlands and deposition of the Eagle Nest formation (Pliocene) tended to destroy the major surficial inequalities. The Eagle Nest formation represents the fillings of a structural basin, whose drainage was not integrated into a through-flowing stream until late Pliocene or early Pleistocene.

Broad valley stage.—In early Pleistocene (?) time drainage from the highlands surrounding the Moreno Valley flowed southward in a single stream which flowed across the divide marking the present limit of the valley and into the drainage basin of Coyote Creek, which today drains past Mora and into the Canadian River (Fig. 1; Pl. 2).

LEGEND
Recent alluvium
Pediment 3
Pediment 2
Pediment 1
Broad Valley Stage
105°15'
Iron Mtn.
Willow Cr. Fan
Eagle Nest
Scully Mtn.
Comanche Cr.
STATE HWY. NO.
Moreno Creek
Miles
N
8000
6250
6500
Jc. N. & S. Forks Moreno Cr.
Elizabethtown
Jc. Comanche & Moreno Cr.
Scully Mtn. - 9765' El.
Projected profile Willow Creek Fan
A
B
C
D

ORED STREAM PROFILES WHICH ARE GRADED

D – PEDIMENT 3
C – PEDIMENT 2
B – PEDIMENT 1
A – BROAD VALLEY STAGE

Projected profi
Willow Creek F

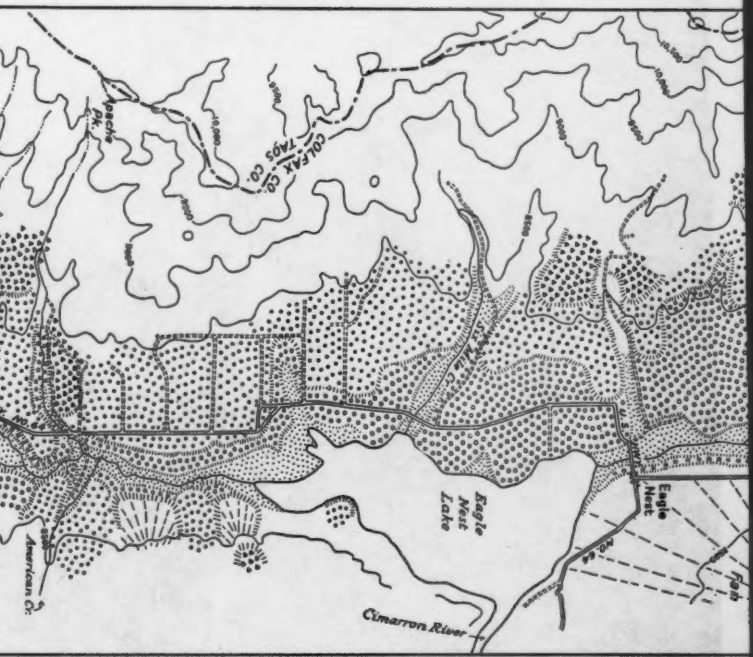
C

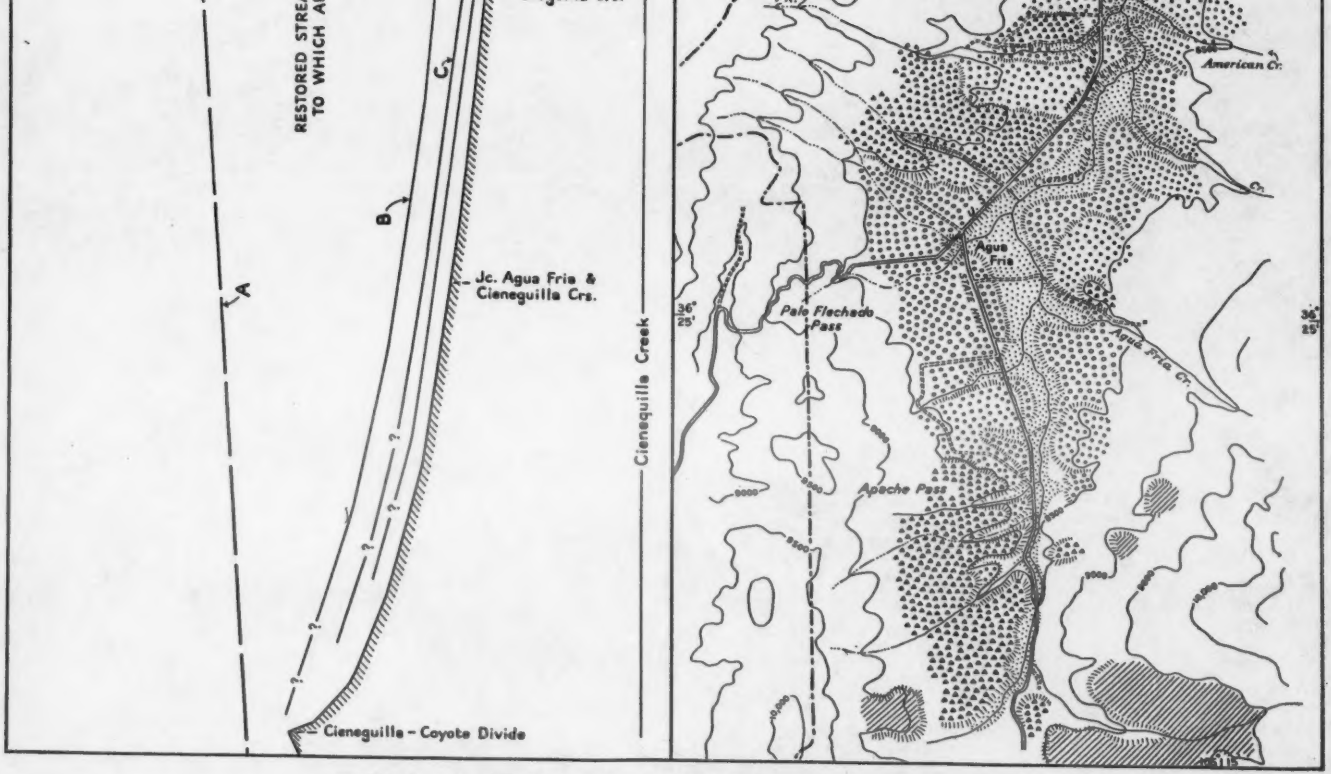
D

Jr. American & Cienequilla Crs.

Eagle Nest Lake

Cimarron River





PHYSIOGRAPHIC MAP AND PROFILES OF MORENO VALLEY

At this period the streams were graded to the surface of the Eagle Nest formation. The grade of the major stream was probably stabilized for a long time by the bedrock divide which now separates the drainage basins of Coyote and Cieneguilla creeks (Fig. 1; Pl. 2). Lateral erosion by tributary streams reduced the interfluvial areas to gently graded surfaces sloping southward and toward the main drainage stream. Continued reduction of the interfluvial areas along the valley walls resulted in a widening of the valley and in surfaces cut on deformed bedrock and graded to the unconsolidated valley fill. In this manner a pediment surface, here called the Broad Valley Stage, was developed and adjusted to a temporarily stabilized base level in a manner similar to those pediments described by Bryan and McCann (1936, p. 150) along the Upper Rio Puerco in New Mexico.

Well-developed bedrock spurs testify to the presence of the old Broad Valley Stage (Pl. 2). Plotting of the remnants shows a gentle southward gradient of approximately 10 feet per mile. The best preserved remnant of this stage is on the eastern valley wall, adjacent to the southern divide (Pl. 2), where an extensive surface is cut across deformed sedimentary rocks belonging to the upper red beds of the Magdalena formation. This remnant is gently graded toward the present valley, and small streams flow across it in wide shallow channels which show no signs of arroyo cutting. The gently sloping banks of the channelways are grass covered, and no exposures of gravel were seen. A deep soil mantles this surface, above which rise highlands capped by basaltic lava flows. These flows, however, do not rest on the surface of the Broad Valley Stage, but on a surface of low relief almost 200 feet above the pediment. They are believed to rest on the Mid-Tertiary erosion surface of low relief and to have been extruded before the faulting which produced the Moreno Valley.

Streams descending from the pediment remnant to the present valley drainage are intrenched in steep canyons. The headward erosion of the streams and lateral widening of the canyons are rapidly reducing the surface area of remnants of the Broad Valley Stage.

Piracy of drainage by the Cimarron River.—The development of the pediment of the Broad Valley Stage was interrupted by a period of stream incision and an accompanying major drainage change. A small stream working headward from the Great Plains cut through the eastern wall of the Moreno Valley, tapping the drainage of the northern portion of the valley and diverting it from its normal southward course into the steep Cimarron Canyon. The steeper gradient of the Cimarron River and its increased volume permitted such rapid downcutting that the southward

drainage of the southern portion of the Moreno Valley was captured and reversed so that it joined Moreno Creek draining from the north. The combined drainage flowed into the Cimarron Canyon. At this time the Moreno Valley assumed its present drainage basin (Pl. 2).

This piracy by the Cimarron River may have been facilitated by a zone of faulting. In part, the Cimarron Canyon may follow a fault zone marking the northern boundary of the Cimarron fault block. At the entrance to the Cimarron Canyon from the Moreno Valley a shift of the river a few hundred feet to the north would have permitted it to have cut through unconsolidated valley fill and Mesozoic strata instead of pre-Cambrian granite gneiss.

All later physiographic development of the Moreno Valley is associated with successive lowering of the stream gradients after this piracy. The Cimarron River thus became the dominant factor in the development of the successively lower pediments of the Moreno Valley, and the Cimarron Canyon became the sole avenue for removal of debris from the valley.

Pleistocene pediments.—In the Moreno Valley three well-developed pediments lie between remnants of the Broad Valley Stage and the flood plain of the present stream. These pediments are cut largely on the unconsolidated valley fill (Eagle Nest formation) and are capped by thick deposits of locally derived subangular gravel and slope wash. These three distinct pediments were formed during periods of stabilization of the successively lowered local base level. The processes operating in the Moreno Valley were similar to pediment-forming processes described by Bryan and McCann (1936, p. 151-152).

Plate 2 shows the position of the successive local base levels to which the pediments of the Moreno Valley are graded. These stream gradients have been reconstructed by restoring the contour lines on the pediment remnants and by projecting the slope of the pediments.

In general, Pediment 1 was graded to a stream about 140-150 feet above the present grade of Moreno and Cieneguilla creeks. The pediment surface slopes upward with a gradient of 2 degrees near the center of the valley to more than 5 degrees near the valley walls. Remnants of this pediment are particularly well preserved in the southern portion of the valley (Pl. 2; Pl. 5, fig. 1).

Formation of Pediment 1 was followed by a period of stream incision and stabilization when the major streams deepened their channels by 75 feet or more to a level approximately 60± feet above the present stream gradients. Pediment 2 (Pl. 2), whose surface approximately parallels that of Pediment 1, was developed during this second period of stream

FIGURE 1. PRE-CAMBRIAN QUARTZITE
Near mouth of American Creek, showing curved and slickensided fault surfaces.

FIGURE 2. CROSS-BEDDED, ARKOSIC, AND CONGLOMERATIC SANDSTONE
Magdalena formation, 4 miles south of Agua Fria.

FIGURE 3. BASAL CONGLOMERATE OF RATON FORMATION
In gully cut into Pediment 2, 1½ miles north of Elizabethtown.

FIGURE 1. PICURIS (?) FORMATION
In gully at Mills Divide. Note hammer for scale. Small fault, possibly due to slumping, in foreground.

FIGURE 2. EAGLE NEST FORMATION
Exposed in old placer workings east of Elizabethtown. Note surface wash, partially cemented by caliche.

PICURIS (?) AND EAGLE NEST FORMATIONS

stabilization. Like Pediment 1, its surface is best preserved along the western valley wall. North of Elizabethtown a large remnant bevels deformed pre-Eagle Nest formations (Pl. 5, fig. 2).

A third period of stream incision followed by stabilization produced the lowest pediment, Pediment 3. This pediment is also parallel to those previously formed. At its lowest point it lies only 20-30 feet above the present stream. In general, Pediment 3 is not extensive, except near Eagle Nest Lake and Agua Fria (Pl. 2), where it covers an area of several square miles. At no point were good vertical sections observed, and it has not been definitely proved that it bevels pre-Eagle Nest formations. However, the slopes, the character of the observed gravel covering, and the general dissimilarity of this surface to the lower flood plain have suggested that it is a pediment and not an alluvial terrace.

Recently Pediment 3 has been dissected and a broad alluvial flood plain formed along the streams (Pl. 5, fig. 1). Throughout most of the valley the flood plains are now being dissected by arroyo cutting.

In general the three lower pediment surfaces are parallel, except at the northern and southern ends of the valley where they tend to spread apart (Pl. 2). Constriction of the Moreno Valley between Scully and Iron mountains has not appreciably interrupted the continuity of the stream profiles to which the pediments are graded. At the time of the earliest pediment, the Broad Valley Stage, Scully Mountain rose less than 600 feet above the major stream drainage. Lowering of stream gradients excavated the surrounding country, and the relative relief was successively increased so that Scully Mountain rose more than 1200 feet above the stream to which Pediment 1 was graded. At present the summit of the mountain lies almost 1400 feet above the level of Moreno Creek.

Piracy of Comanche Creek.—When Pediment 1 constituted the floor of the Moreno Valley, Comanche Creek (Pl. 2) flowed from the mountains to the axis of the valley along the north side of Scully Mountain. A small stream working headward between the western margin of Scully Mountain and the valley wall was carving a broad headwater basin. Because of steeper gradient and favorable bedrock conditions—for the stream was cutting parallel to the strike of the formations—this small creek proved to be more active than Comanche Creek during the period of stream incision after the completion of Pediment 1 and previous to the development of Pediment 2. In this interval the small stream broke through the headwall of its drainage basin and captured the drainage of Comanche Creek where it issued from the mountain canyon.

Diversion of the upper drainage area of Comanche Creek has so increased the activity of this small stream that it has been able to maintain itself during the remainder of the physiographic development of the Moreno Valley. All traces of the former eastward course of Comanche Creek have been obliterated.

Alluvial fan of Willow Creek.—North of Eagle Nest Lake and east of Moreno Creek lies the great alluvial fan of Willow Creek (Pl. 2). This fan, whose average slope is about 3° to the southwest, completely covers the lower pediment remnants which existed in this area, with the exception of a small marginal remnant of Pediment 3 near the flood plain of Moreno Creek.

The drainage basin of Willow Creek lies entirely in the area of Mesozoic and Tertiary shales and sandstones. Stream action has produced so much debris because of the relative incompetency of these rocks that Willow Creek has been choked and fan material has been deposited where the steep stream gradient of the mountains is replaced by the gentler gradient of the valley floor.

Within the past 60 years placer mining in the upper canyon section of Willow Creek has spread large quantities of alluvial material on the fan. Present inactivity of placer mining has allowed vegetation to cover and conceal much of this recent accumulation.

Age and correlation of the pediments.—The earliest pediment, the Broad Valley Stage, is believed to be Early Pleistocene, although it may possibly have been initiated in Late Pliocene time. It was developed by streams graded to the surface of the Eagle Nest (Pliocene?) formation. Pediments 1, 2, and 3 are assigned to the Pleistocene. Actual dating of these surfaces can be accomplished only after the physiographic history of the Great Plains is known. The later physiographic history of the Moreno Valley was dependent on the changes in gradient of the Cimarron River, which in turn reflect the changes taking place in the Plains province. The history of that region is complicated by the long-continued volcanic activity. No attempt has been made to correlate the Quaternary history of the Great Plains and the Moreno Valley.

Contrary to published reports (Ellis, 1931; 1935) the Moreno Valley has not been glaciated (Ray, 1940). Although some of the streams draining into the Moreno Valley may have originated in more or less permanent Pleistocene snow masses in the high mountains, it is certain that no glacial debris was ever carried into the valley. Glaciers which are known to have existed in the mountains to the west of the Moreno Valley all lay within the drainage of the Rio Grande.

CENOZOIC HISTORY OF THE MORENO VALLEY AND ADJACENT REGIONS

The Moreno Valley has been affected by the major geological events which have produced the structurally and topographically complex Southern Rocky Mountains of New Mexico and southern Colorado. An attempt is made to correlate the events in the Moreno Valley with those of adjacent regions. Recent geologic investigations in the nearby Rio Grande Depression have been summarized by Bryan (1938), so that no attempt is here made to review the rapidly accumulating literature of that important region, which is somewhat similar to the Moreno valley. The writers have had access to much unpublished material and wish to acknowledge their indebtedness to the many recent workers in the area (Bryan, 1940), with whom, over a period of years, they have had many discussions, both in the field and laboratory. In numerous instances the tentative suggestions and correlations here made disagree with both published and unpublished information.

Laramide folding, in some regions extremely complex (Burbank and Goddard, 1937), and upwarping of the Southern Rocky Mountain province as a broad geanticlinal structure ended the shallow Late Cretaceous marine deposition. This geanticlinal structure in New Mexico extended from the Great Plains on the east, across the Sangre de Cristo Mountains, the Rio Grande Depression, and the Conejos and Jemez extensions of the San Juan Mountains, to die out where the mountains merge with the Colorado Plateau on the west.

There ensued a long erosional interval during which coarse clastic sediments of the Raton formation (Eocene) were deposited in the region of the Moreno Valley. These are believed to have been derived from a highland to the west, for, as noted by Lee (1922, p. 7), the basal conglomerate of the Raton is thicker and coarser to the west. In the Moreno Valley region relief may not have been great at this time. In the San Juan Mountains, 150 miles to the northwest, Atwood (1940, p. 300) has postulated great heights, possibly as much as 12,000 to 15,000 feet above sea level, to explain the occurrence of a till assigned to the Eocene which indicates alpine glaciation at a time of mild climate.

In post-Raton time and previous to the deposition of the Picuris (?) formation, there was in the Moreno Valley region a period of igneous intrusion. Rhyolite and monzonite porphyry cut the Raton and all older formations and in the region of Scully Mountain warped and deformed the strata. This igneous activity is believed to be related to that of the Spanish Peaks region to the north (Hills, 1901; Knopf, 1936).

The long interval between the deposition of the Raton and Picuris (?) formations in the Moreno Valley has left no records which give a

clue to the geological events of that time. Whether there was deposition followed by erosion of Early to Mid-Tertiary sediments is a matter of speculation. Future studies in the High Plains may ultimately reveal a sequence of events for which data in the mountains is lacking. Unpublished studies by Ray (1938) in the northern Colorado Front Range indicate a post-Oligocene uplift of the mountains relative to the plains and the production in Miocene time of a surface of erosion in the mountains continuous with a surface of deposition in the plains. A similar Mid-Tertiary erosion surface is believed to have been developed in the Sangre de Cristo Mountains of southern Colorado and New Mexico.

In the vicinity of the Moreno Valley numerous remnants of an erosion surface occur at elevations of 10,000 to 11,000 feet. A careful examination of the topographic map (U. S. Geol. Survey, Taos quadrangle, Advance Sheet, 1938) shows numerous flat-topped spurs and areas of low relief rising above the present stream valleys. Sawmill Park (11,000+ feet), Bobcat Park (10,000± feet), Cabresto Park (10,800± feet), Garcia Park (10,000± feet), and many other areas of low relief attest the presence of an old erosion surface (Pl. 6, figs. 1, 2). To the north, in the vicinity of the Spanish Peaks, a similar surface, possibly of the same age, rises to a height of nearly 10,000 feet on the east side of the mountains (Knopf, 1936, p. 1729) and to higher elevations on the west. The writers believe that the San Juan peneplain was continuous with the erosion surface of the Sangre de Cristo Mountains and the depositional surface of the Great Plains. Atwood and Mather (1932, p. 25) state that peneplanation occurred during Late Pliocene time because it "cuts across volcanic strata . . . which are classified as Miocene." Atwood (1940, p. 305), however, has classified the San Juan peneplain as Mid-Tertiary. An earlier erosion surface in the San Juan Mountains, as yet little known, should be mentioned as a possible correlative of the erosion surface in the Moreno Valley region. The "San Juan Peneplain" of Atwood and Mather may not be represented in this area. Present data are not conclusive. Atwood (1940, p. 305-308), however, correlated the Mid-Tertiary surface of low relief (San Juan Peneplain) with many of the erosional surfaces described throughout the Southern Rocky Mountains. Studies by Ray suggest that many of the so-called peneplains are complex and belong to several different periods. This interpretation may also be true in the San Juan Mountains. Some of the complexities of this problem may be gained from the studies of Van Tuyl and Lovering (1935; 1936) and Rich (1936). Nevertheless, the writers believe that by the close of Mid-Tertiary time a broad and featureless plain of deposition had developed to the east of the mountains which was continuous with an erosional surface in the mountains.

FIGURE 1. SOUTHERN PORTION, LOOKING NORTH
Pediment 1 in background; in foreground flood plain of Cieneguilla Creek, slightly dissected by arroyo cutting.

FIGURE 2. PEDIMENT 2 NORTH OF ELIZABETHTOWN
Bevels deformed strata and igneous intrusives. Moreno Creek on right.

FIGURE 1. LOOKING SOUTHEAST FROM WESTERN VALLEY WALL
Eagle Nest Lake in midground, Great Plains in background, beyond Cimarron Range. Flats in right background (indicated by arrow) are remnants of Broad Valley Stage.

FIGURE 2. NORTHERN PORTION, LOOKING NORTHWEST FROM WESTERN SLOPE OF BALDY MOUNTAIN
High flat areas belong to Mid-Tertiary erosion surface. High peaks of Sangre de Cristo Mountains in background.

In part, the erosion surface was one of low relief, and in part a surface of maturity. Monadnocks and groups of rugged mountains rose to elevations of several thousand feet above the general level. In northern New Mexico there was no Rio Grande Depression and no Moreno Valley. Although definite evidence is lacking, it is possible that the featureless surface in the Great Plains was continuous with the erosional surface of the Sangre de Cristo Mountains, and with the broad erosion surface in the San Juan Mountains (the San Juan Peneplain?).

A remnant of the deposits of debris which were spread on the erosional surface occurs at Mills Divide in the Moreno Valley and has been tentatively correlated with the Picuris formation. This deposit was made by streams flowing across the old Mid-Tertiary surface before its deformation.

Large portions of the Mid-Tertiary surface were covered with lava flows and volcanic ejecta, as in the Raton Mesa region and adjacent to the southern part of the Moreno Valley, and possibly in the Rio Grande Depression. Volcanic activity has persisted throughout this region until very recent times.

Profound faulting and uplift about the close of Miocene time or very early in Pliocene time destroyed the Mid-Tertiary landscape of low relief. The Laramide geanticline of the Southern Rocky Mountains was uplifted and invaded by faulting similar in character to that of the Basin and Range Province, resulting in the production of a series of downfaulted basins. The Rio Grande Depression was initiated at this time as a series of basins, whose general structure extends to the north as the San Luis Valley, South, Middle, and North Parks of Colorado. The Moreno Valley was developed at this time as a small basin, similar in structure and trend to the great Rio Grande Depression.

Following the faulting, uplift, and the development of the Moreno Valley, locally derived debris accumulated in the valley as the Eagle Nest (Pliocene?) formation. Integration of the drainage produced a southward-flowing stream to which were graded pediments which have been named the Broad Valley Stage.

The Rio Grande Depression had a more complicated history. Debris carried into the isolated basins of the depression are interbedded with lava flows. A complex series of erosional surfaces was developed below the Mid-Tertiary surface by slight changes in stream gradients, caused by continued movement along the faults, alluviation of the basins, and integration of the drainage, or by periodic volcanic outbursts. Correlations and the complete history in the Rio Grande Depression are still far from completely known, despite the active field work under way in

this region (Bryan, 1940). Final correlations must await more detailed data.

In the northern Colorado Front Range differential uplift of the mountains relative to the plains and faulting warped the Miocene erosional-depositional surface. In the mountains the Mid-Tertiary surface was partially eroded, and debris was carried eastward and deposited far from the mountains as the Ogallala (Pliocene) formation. Near the mountain front deformed Miocene sediments were bevelled by a Pliocene erosion surface which is continuous with the depositional surface of the Ogallala formation. It is this compound surface which is the High Plains.

A similar sequence of events must have taken place in northern New Mexico, and it is believed that the stream to which the Broad Valley Stage of the Moreno Valley was graded must have been contemporaneous with the High Plains surface. Slight movements have continued along the faults throughout this region into Quaternary time. The Pliocene(?) Eagle Nest formation is faulted and deformed.

Piracy by the Cimarron River has related the development of the lower pediments of the Moreno Valley to the complex succession of recent events in the upper drainage basin of the Canadian River, where volcanic activity has not been studied by the writers, and no attempt is made to correlate its intricate history with events occurring in the Moreno Valley after the development of the Broad Valley Stage.

SUMMARY OF CONCLUSIONS

This study shows that the Moreno Valley is a structural basin whose location has been largely determined by bounding faults along the east side. Genetically it is in certain respects similar to the Rio Grande Depression about 15 miles to the west, on the western side of the Sangre de Cristo Mountains. The Moreno Valley is located along the complex eastern structural boundary of the Sangre de Cristo Mountains which are part of a geanticline produced by Laramide orogeny. Later Tertiary intrusion, deformation, erosion, and deposition have modified this geanticline and produced the present heterogeneous landscape.

The Moreno Valley was initiated about the close of Mid-Tertiary time by faulting, probably contemporaneously with the Rio Grande Depression. A tentative correlation of geologic events in the development of the Moreno Valley with the regional sequence of events has been attempted. In the description of its physiographic development during the Quaternary, the Moreno Valley has been treated as a unit. No attempt has been made to correlate events in its physiographic history with events in the Great Plains Province.

Because of its relative isolation and the lack of topographic maps and airplane photographs, much of the country to the south and east of the Moreno Valley has been untouched by geological investigators. Within these areas may lie the answer to some of the unsolved problems concerning the complex geologic history, whose whole story cannot be gathered from any single small unit, but which will be told only by the fitting together of the widely spread bits of information gathered by the many workers throughout this region. Gaps which occur in the geologic history of the Moreno Valley may be filled in by studies in nearby areas, and these in turn will fall into a broad regional pattern of geologic events.

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Michigan State College, East Lansing, Mich.; Agricultural and Mechanical College of Texas, College Station, Texas.

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ORIGIN OF LINE OF COLOR CHANGE IN RED BED DEPOSITION

BY G. E. ANDERSON *

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ABSTRACT

The line of color change in Pennsylvanian and Permian sediments, extending from south-central Kansas across Oklahoma into north-central Texas, is produced by a facies change in which a nonmarine and marine series of sediments were deposited contemporaneously during a differential downwarp that changed their initial strike. Later truncation to the present erosional surface has cut the facies change and gives rise to the seeming anomaly that the line of color change transects the present strike of the sediments.

INTRODUCTION

The red beds of Oklahoma occupy approximately the west half of the State. They strike north-south and dip westerly. The dip ranges from more than 50 feet per mile in the Arbuckle Mountain region to less than 30 feet per mile on the Kansas-Oklahoma State line. In the west they dip under the Cretaceous and later sediments, and on the east they are bordered by marine light-colored sandstones, shales, and limestones which dip under the red beds. Along the strike northward into Kansas and southward into Baylor and Archer counties, Texas, the red beds grade into marine sediments with limestones increasing in abundance.

The strip involving the change from red to light-colored marine sediments is known as the line of color change. It forms a zone in the out-

* Deceased August 31, 1940. Paper revised by C. E. Decker and William Ham in response to editorial suggestions.

crop of 25 to 30 feet of interstratified red and marine sediments sharply contrasted between superjacent red bed and subjacent light-colored strata. There is no apparent stratigraphic break in strike and dip between the red beds and marine sediments.

The sediments in the area of color change in Oklahoma are Upper Pennsylvanian (Pontotoc) and Lower Permian (Stillwater, Wellington, Garber, and Hennessey). In Texas these are equivalent to the Upper Cisco, Wichita, and Clear Fork (Sellards, Adkins, and Plummer, 1932). In Kansas they range from the Howard limestone (Beede, 1909), Upper Pennsylvanian, through the Lower Permian to the equivalent of the Hennessey shale.

The line of color change emerges from under the post-Permian sediments in southeast Reno County, Kansas, (Fig. 1) zigzagging diagonally across the strike of the sediments and continuing in a southeasterly course to Payne County, Oklahoma. Then it trends southward, changing in Seminole County to a southwesterly course which continues to the north border of the Arbuckle Mountains. South of these Mountains the color line is covered by basal sandstones of the Lower Cretaceous from which it emerges in southwestern Montague County, Texas and continues westerly through Clay, Archer, and Throckmorton counties.

The line of color change cuts across the Permian-Pennsylvanian boundary in northern Payne County, Oklahoma, and continues in Pennsylvanian (basal Pontotoc) sediments parallel to the Pennsylvanian-Permian boundary to Archer County, Texas, where it again crosses this boundary and continues to the southwest transecting the strike of the Lower Permian sediments. By transecting the strikes, the line of color change forms a zigzag pattern of red bed tongues projecting from the central red bed area northward and southward into the marine sediments of Kansas and Texas.

The red beds, from their base to near the top of the Hennessey shale, were deposited in a nonmarine environment, as indicated by the presence of Amphibia, many reptiles, and land plants (Romer, 1935). The land flora and fauna and the striking red color of the red bed tongues are used to trace them into the marine sections of Kansas and Texas. Alternating with these are fossiliferous marine tongues projecting into Oklahoma southward from Kansas and northward from Texas. The latter are indicated by marine limestones extending well into the red bed area.

The strike of the sediments, passing from the nonmarine red bed facies into the marine facies across the line of color change, is not satisfactorily explained by simple lateral gradation along the present

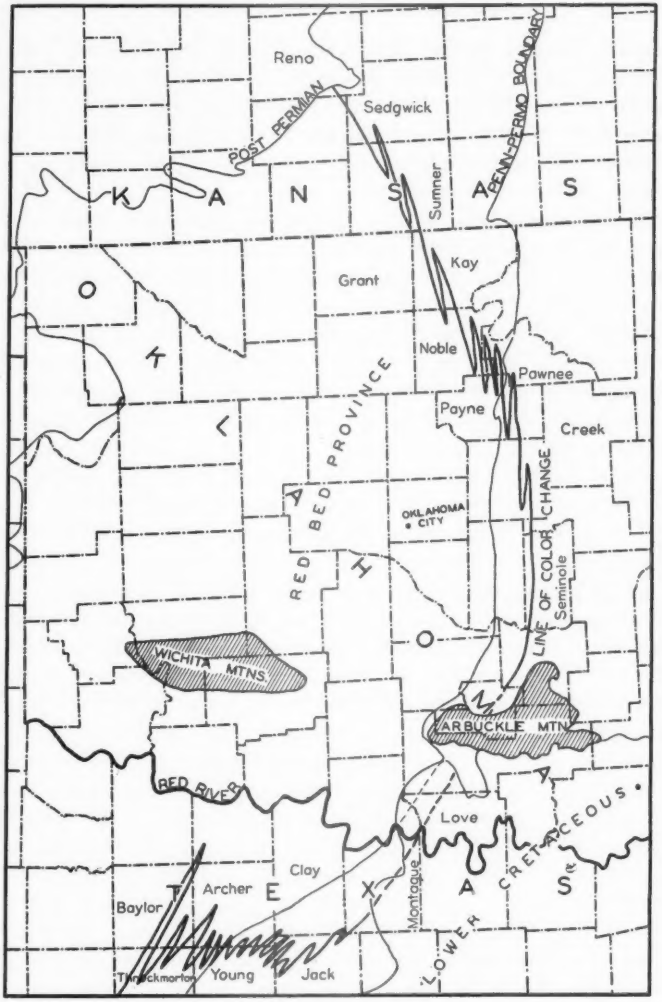

FIGURE 1.—*Index map showing line of color change in Oklahoma, Kansas, and Texas*

strike, since lateral gradation must have the same depositional environment. In this case the line of color change passes from a marine environment to one of subaerial deposition.

THEORY OF THE COLOR CHANGE

Clastic sediments eroded from a geanticline are deposited contemporaneously over the adjacent piedmont, the alluvial plain areas, and in the epeiric sea beyond. In an arid climate the piedmont and alluvial deposits form a series of red sandstones and shales, deposited contemporaneously with a marine series of light-colored sandstones and shales in the adjacent epeiric sea.

Numerous oscillations of the epeiric sea cause the shore line to advance and recede over the alluvial plain and outer piedmont. Alternating tongues of red beds and marine sediments are thus deposited within the area of the oscillating shore line; the former project into the marine sediments basinward and the latter extend into the red beds landward. Thus, in a section through the marine and red bed sediments taken at right angles to the shore line, the contact between the alternating red and marine strata forms an ascending zigzag line, constituting the line of color change. The latter, therefore, represents a facies change along the shore line separating the adjacent depositional environments of the alluvial plain from the epeiric sea. In Figure 2 the line of color change is the zigzag line forming the contact between the marine and red sediments within ABCD. The dip is exaggerated, for the sediments would normally have a slight initial dip toward the marine sediments at the angle of the seaward slope of the alluvial coastal plain. The shoreward edge or apex of the marine, and seaward edge of the red bed tongues, coincide approximately with the shore line and, hence, are parallel with the initial strike of the sediments.

If a horizontal section were taken at AB, A′B′, and DC respectively (Fig. 2), the section at AB would truncate only red bed sediments, and section DC, only the marine sediments. In the truncated section A′B′, both red and marine sediments are exposed, and the color line coincides with the initial strike of the sediments. If the foreground of the block were depressed by rotation so that the line AB approximates the position formerly occupied by DC, and if the block were truncated by a horizontal erosion plane (Fig. 3), this plane would cut marine sediments in the background and red beds in the foreground. The region between the truncated marine sediments and red beds would consist of alternating marine tongues projecting into red sediments, as they occur (Fig. 1) in Sedgwick and Sumner counties, Kansas; Kay, Noble, Pawnee, and Payne counties, Oklahoma; and Clay, Archer, Baylor, Throckmorton, Young, and Jack counties, Texas.

As the foreground is depressed, the outcrop migrates up the dip, correspondingly changing the direction of the initial strike of the

sediments from north to northwest. Since the initial strike coincides with the apices of the interstratified tongues of red bed and marine sediments, the strike of the sediments, after subsidence and truncation, cuts the line of color change at some angle depending on the amount of downwarp.

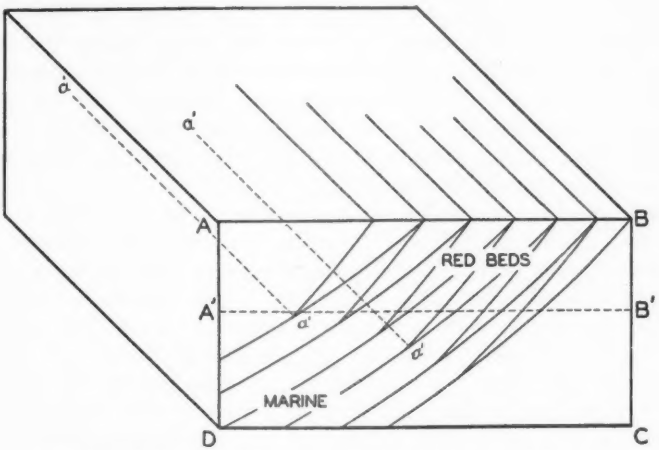

FIGURE 2.—*Block diagram of contemporaneous marine and red bed deposition*

APPLICATION OF THE HYPOTHESIS TO THE OKLAHOMA RED BEDS

During the late Pennsylvanian and early Permian, central Oklahoma was covered by an epeiric sea. Its fluctuating eastern limits extended northeasterly over the area of the present line of color change. East of the shore was a comparatively wide alluvial plain and piedmont area, and the source region of a major volume of these clastic sediments lay farther eastward. During the Permian, evaporites were deposited which included rock salt in the Stillwater, Wellington, Garber, and Hennessey. In the last three formations the thickest evaporite section is in Woodward County, Oklahoma, suggesting the location of the axis of the Permian basin. The position of the eastern margin of the sea approximates the region of contact between the marine and nonmarine strata. The westward limitation of the individual tongues of the red bed series, as for example a′a′ (Fig. 2), would determine that shore line and, similarly, the easterly extent is determined by the extension of marine tongues into the red beds.

The clastic sediments, partly derived from northern Llanoria, were spread over the piedmont and alluvial plain as red beds on the eastern shore of the epeiric sea (Van der Gracht, 1931; Miser, 1934; Green, 1936). Some sediments were carried to the west into the adjacent sea. Accordingly, two series of sediments were deposited contemporaneously, shale and sandstone on the alluvial plain and the lower piedmont, and a marine series of light-colored sandstones, bluish and gray shales, some limestones, and evaporites in the sea to the west. At the shore line, these were interstratified, due to the oscillation of the shore, resulting in the intertonguing of the marine and red bed series. During submergence marine sand and shale were deposited upon nonmarine shale, during emergence nonmarine red shale and sand were deposited on the marine sandstone.

Had deposition of the intertonguing series taken place without differential subsidence during or following deposition, the strike of the sediments would parallel the apices of the marine and red bed tongues. During deposition, however, a downwarp in the Oklahoma area permitted the accumulation of a greater thickness of red beds than in the areas of Kansas or Texas. The downwarp hinged east of the shore line and reached a maximum in the axis area of the Permian basin of western Oklahoma. Deposition kept pace with the downwarp, preventing the invasion of the sea to the east and maintaining the shore line in approximately its northeast-southwest direction. The absence of marine tongues through the central area of the red beds indicates this condition. Fifteen thousand feet of Permian and Pennsylvanian sediments filled this trough. With the uplift of the Arbuckles and Wichita Mountains in a northwesterly-southeasterly direction across the downwarp, the Anadarko basin was formed north of the latter group. Both uplifts contributed local sediments and were apparently independent of the great downwarp.

Truncation has removed whatever red beds were deposited in Kansas and Texas north and south of the line of color change, leaving a thick truncated remnant in the downwarped Oklahoma area. Accordingly, in Kansas and Texas erosion has exposed marine sediments, while in the downwarped area of Oklahoma, the red beds have been preserved. This stage is represented diagrammatically in Figure 3 for the north end of the red bed area. Because the differential downwarp increases westward toward the central area of the Permian epeiric sea, as well as southward along the shore line in the Oklahoma area, the initial strike of the sediments in the line of color change has been changed to a general northwesterly direction in Oklahoma and a southwesterly

direction in Texas. Along the bedding planes basinward to the west, the facies change from red beds to marine has been confirmed by core drilling in Oklahoma, and is known to exist in Texas (Sellards, Adkins, and Plummer, 1932).

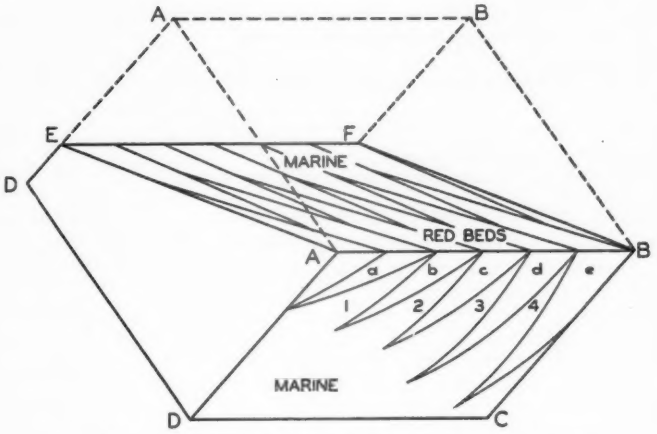

FIGURE 3.—*Block diagram of area after distortion and truncation*

SUMMARY

According to this interpretation of the line of color change, the oscillating shore line has been reflected by corresponding tongues of marine sediments at the north and south ends of the downwarp within the red beds. The intertonguing resulted from three general processes,—deposition, deformation, and erosion. A series of alternating marine and nonmarine sediments was deposited near an oscillating shore line; the deposits were tilted basinward; and later, erosion truncated the formations exposing the diagonally tilted edges of the interfingering tongues. Because of the contemporaneity of the intertonguing sediments, correlation of the Kansas and Texas marine sections should be possible.

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JURASSIC STRATIGRAPHY OF CENTRAL OREGON

BY R. L. LUPHER

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ABSTRACT

Ten formations are recognized in the Jurassic marine sequence of central Oregon. Angular unconformities separate these formations from the Paleozoic and Upper Triassic formations below and the Upper Cretaceous above. No marked diastrophism is indicated within the Jurassic sequence, though long intervals of nondeposition are present.

A chronologic arrangement of the formations shows four formations in and near the Middle Lias, four in the early Middle Jurassic, one in the early Upper Jurassic, and one in the Upper Jurassic or Cretaceous. A marked threefold lithologic grouping of the sequence is evident but does not correspond to the chronologic arrangement. The oldest lithologic division consists of late Lower Lias shales and sandstones, over 2000 feet thick, at the isolated Donovan Ranch locality in the Silvies River Canyon near Burns. The second lithologic division contains three Lias formations and two early Middle Jurassic formations at the base of the section in the main Jurassic area. It is characterized by thin calcareous formations abounding in well-preserved fossils. The third division includes the remaining part of the Jurassic sequence consisting of two formations of the early Middle Jurassic and one of the early Upper Jurassic, as well as the uppermost formation of uncertain age. It is characterized by thick, noncalcareous formations of sandstone and shale in which fossils are less abundant and not well preserved.

INTRODUCTION

Paleozoic and Mesozoic marine rocks have been known in central Oregon since 1865 when a cavalry expedition under Captain Drake

collected fossils from the upper Crooked River region and gave them to Dr. Thomas Condon (McCornack, 1928). Subsequently a number of writers mentioned either pre-Tertiary marine rocks or fossils in central Oregon. The known occurrences were, however, confined to three restricted areas: (1) the upper Crooked River country and eastward to Izee on the South Fork of the John Day River; (2) the small isolated outcrops of the so-called "Hardgrave Jurassic" in the Silvies canyon, 19 miles north of Burns; and (3) some black shale in the Canyon City region, 30 miles north of the "Hardgrave" locality. That these areas were only marginal tracts of a pre-Tertiary marine area of some 700 square miles was not visualized until 1926 when Dr. E. L. Packard discovered a large area of marine rocks in the upper Crooked River country, the South Fork of the John Day valley, and northeastward across the headwaters of the Silvies River almost to Canyon City in the John Day River area. Subsequent investigations have shown that these rocks include a number of fossiliferous formations representing the Mississippian, Triassic, Jurassic, and Cretaceous periods.

In the summer of 1927 and intermittently since then the writer has studied the Jurassic rocks and their faunas. Reconnaissance soon revealed that Lower, Middle, and Upper Jurassic rocks constituted much of the pre-Tertiary terrane. Fossils are very abundant in comparison to the sparse faunas of other North American Jurassic rocks.

The Jurassic rocks, including plutonics which are possibly of Jurassic age, form a triangular area, about 45 miles on each side, with Canyon City at the northeast corner, Burns at the southeast corner, and Suplee Post Office at the west corner. The principal area of Jurassic marine rocks (Fig. 1) is 12 to 18 miles wide, and extends from the vicinity of Suplee eastward for 40 miles across the upper drainage basins of Crooked River, South Fork of the John Day River, and Silvies River. Other small isolated areas appear south of this main area as far as Harney Basin.

This paper describes the succession, character, and distribution of the Jurassic formations and faunas and presents evidence for the dates assigned to them. The age assignments here proposed are based largely on the ammonites which give more exact results than the other fossils. Even though the ammonite faunas are exceptionally rich and varied, many of the genera and most of the species are new and undescribed. Consequently, the presentation of evidence for age assignments involves the identification of short-ranging families and genera that are known elsewhere, and the comparison of closely related species and genera. This is difficult to present in advance of systematic treatment of the faunas because the very detailed biologic subdivision promoted by

some writers in recent years has caused a wide variance in present concepts of taxonomic groups of Jurassic ammonites. All that is done here is to identify or compare the new Oregon forms with known families, genera, and species that have pertinent chronologic value without attempting to express opinions regarding the validity of the names as used by other writers.

The Jurassic faunas of North America are not sufficiently abundant and well known to permit accurate short-range correlation of most of the Oregon faunas. Therefore, much of the correlation evidence rests upon a comparison of Oregon and European faunas. Since we lack a detailed Jurassic sequence in this country, it will be necessary to define the chronologic and stratigraphic positions of the Oregon faunas and strata in terms of units that are of European origin. Most of the recent correlations of North American Jurassic faunas have been stated in terms of the detailed chronology developed in England by Buckman (1909-1930), and this appears to be legitimate if its limitations are recognized. Therefore, the chronologic positions of the Oregon faunas will be given, insofar as possible, in terms of Buckman's "ages." Corresponding European stratigraphic units will be employed in the sense that they are used in the standard reference work by Arkell (1933).

PREVIOUS WORK

Previous work on the pre-Tertiary rocks of central Oregon before 1926 has been outlined by Packard (1928). Insofar as the Jurassic is concerned, there have been few published statements based upon first-hand information. In 1899 Washburne (1903) crossed a part of the Triassic and Jurassic rock area between Suplee and Canyon City but referred to the rocks as "possibly Triassic" shales and as "basalt." Washburne also stated (p. 227) that Thomas Condon had collected a few invertebrate species, identified by Stanton as possibly of Jurassic age, from "Lower Beaver Creek" in 1862. These probably were collected in 1865 or 1866 (McCornack, 1928, p. 38) from upper Beaver Creek.

Other references to the Jurassic include a paper by Packard (1928) which stated the general distribution of the Jurassic marine rocks and their relationships to other systems; a description of a rudistidlike pelecypod from the Lower Jurassic of the Suplee region (Lupher and Packard, 1930); and two abstracts of progress reports (Lupher, 1930, 1931).

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GENERAL GEOLOGY

The principal topographic and geographic features of the region may be better visualized in connection with some of the general geologic features. The pre-Tertiary rocks lie mostly in the dissected portions of a high, warped plateau south of the Ochoco Range. The plateau is the high northern margin of the southern Oregon lava plain which rises from an altitude of 4000 feet in the Harney Basin to 5400 feet at the base of the Ochoco Range. Major streams radiate from this elevated tract, some reaching the Deschutes, Columbia, and Snake rivers, and some flowing southward into the Great Basin. That the area of pre-Tertiary rocks was once largely covered by Tertiary lavas and continental sedimentary deposits is attested by the sinuous margin of the Tertiary deposits almost completely encircling the area and by the many Tertiary outliers upon it. The origin of the pre-Tertiary exposures cannot be accounted for solely by erosion of a window through a thick lava section such as the Columbia River lavas. The isolated remnants of the Tertiary cover within the old-rock area are mostly Pliocene rhyolite and basalt less than 200 feet thick. The Tertiary cover thickens rapidly on all sides and includes Miocene and possibly earlier rocks. Apparently the pre-Tertiary area stood as an elevated tract at the south margin of the Columbia basalt floods and remained high while the surrounding areas were built up by Miocene and Pliocene sediments and lavas. The degradation of the elevated tract and the filling of the surrounding areas came near a common level during the Pliocene, and thin flows of rhyolite and basalt spread over much of the former highland and lowland regions, completing what has been termed the "Silvies Surface" (Lupher, 1937). Subsequent

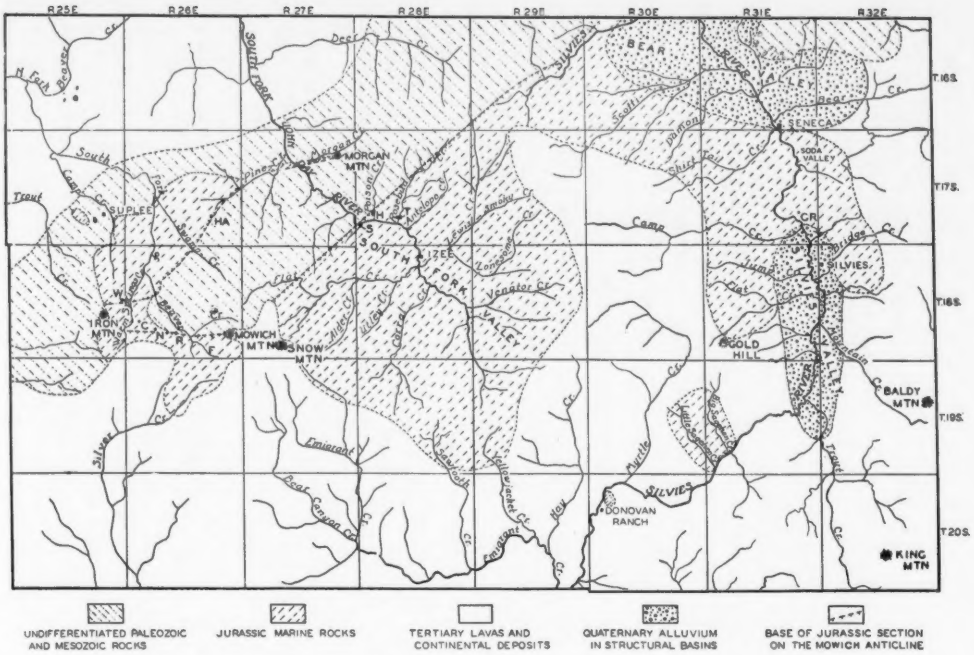

FIGURE 1.—*General distribution of Jurassic marine rocks in central Oregon*

The map is too generalized to show details such as the irregularities of the lava margin and the many Tertiary outliers upon the pre-Tertiary area. C, Colpitts Ranch; CR, Craddock Ranch; F, Freeman Creek; H, Hyde Ranch; HA, Harris Ranch; N, Nicely Ranch; R, Robertson Ranch; S, South Fork Bridge; T, Trowbridge Ranch; W, Weberg Ranch.

elevation, warping, and faulting initiated the present erosion cycle and caused the removal of extensive areas of the thin Tertiary cover.

The deformation of the surface outlined some of the principal topographic features of the region (Fig. 1), prominent among which are the

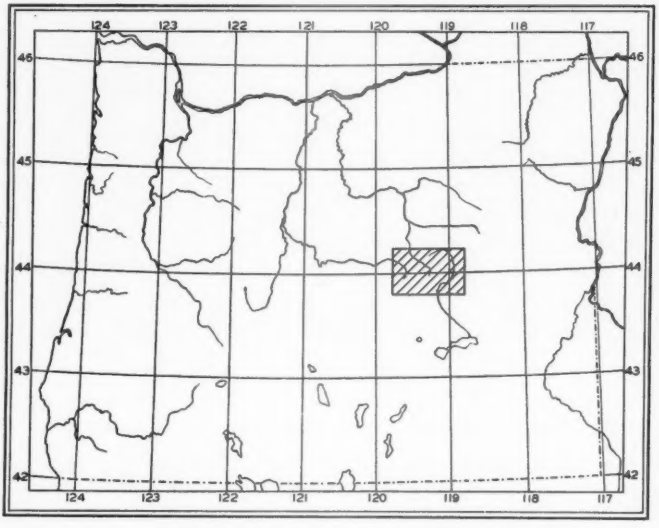

FIGURE 2.—*Outline map of Oregon*
Showing location of area covered by Figure 1.

downward basins. Bear Valley and Silvies Valley are alluviated structural basins lying from 400 to 600 feet below the general level of the plateau. Between these basins the upper canyon of the Silvies River (Pl. 4, fig. 2) notches into the plateau, and southward, between Silvies Valley and the much broader downwarp of the Harney Basin, the deep lower canyon of the Silvies trenches the plateau for a distance of 27 miles. Jurassic strata are exposed on the eroded margins of the Silvies and Bear Valley downwarps and in the upper canyon between them. Though the lower canyon is about 800 feet deep in places, it cuts into the thick Tertiary section that encircles the old rock area and so exposes Jurassic rocks only where it cuts across buried hills and elevated tracts of the ancient topography.

The South Fork of the John Day River also heads in a structural basin that is separated from the Silvies Valley to the east by a narrow north-

ward-projecting peninsula of the plateau lava. Heading in the downwarp and flowing northwest and north across the Ochoco Range on a steep gradient, the upper South Fork has greatly enlarged the structural basin into the deeply dissected "South Fork Valley" (Pl. 2, fig. 2) which is about 16 miles wide. Only remnants of the Tertiary cover remain in the valley, and the Jurassic and Triassic rocks are widely exposed here and in the upper courses of tributary streams that notch deeply back into the lava rim.

The escarpment of the southward-tilted fault block of Snow Mountain (Pl. 1, fig. 1) rises abruptly south of the valley to an altitude of over 7000 feet and forms the three-way divide between John Day River, Crooked River, and the Great Basin. The Tertiary cover is so thin in some places south and southeast of the South Fork Valley that isolated areas of Jurassic rocks are exposed along some western tributaries of the lower Silvies River, such as Big Sagehen Creek, Little Sagehen Creek, Hay Creek, Sawtooth Creek, and Latigo Creek. West of the South Fork the plateau slopes westward, and the tributaries of the Crooked River have extended the Jurassic exposures westward beyond Suplee Post Office (Pl. 1, fig. 2).

GENERAL CHARACTER OF THE JURASSIC SEQUENCE

All the rocks in central Oregon of known Jurassic age are of sedimentary character, but at least two sets of intrusives may be of Jurassic age. Basic plutonics, now largely altered to serpentine, were intruded into Upper Triassic marine rocks along the Ochoco Range and southward as far as Deer Creek, but they were uncovered by erosion before the deposition of the Lower Jurassic Mowich group. The gabbro and diabase mass of Canyon Mountain, reported by Lindgren (1901), and similar rocks on Fields Peak intrude sedimentary rocks which are thought to be of Triassic age, but these intrusives are some distance north of known Jurassic exposures. A third set of intrusives, consisting of numerous dikes and sills of green and black basalt porphyry, was intruded into the Jurassic and other pre-Tertiary rocks. They post-date all the Jurassic marine sequence and closely resemble intrusives in the Tertiary Clarno formation of the Mitchell region.

The Donovan formation appears to be the oldest Jurassic formation in central Oregon, though its actual stratigraphic relations to other formations can only be inferred because of its isolated occurrence in the Lower Canyon of the Silvies River. The stratigraphic order of all the other units shown in Table 1 has been determined from the superposition of these units in the main Jurassic area. No single section shows all the

FIGURE 1. SNOW MOUNTAIN (LEFT) AND MOWICH MOUNTAIN
View southeastward from Beaver Creek-Warm Springs Creek divide showing southward-tilted lavas on skyline and characteristic exposures of Triassic and Jurassic formations below.

FIGURE 2. VIEW NORTHWEST FROM SNOW MOUNTAIN
Looking down upon dissected plateau area in western end of pre-Tertiary rock area.

SUPLEE-SNOW MOUNTAIN AREA

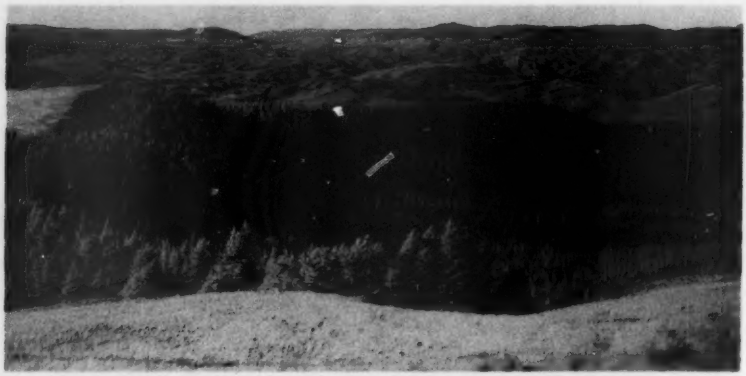

FIGURE 1. VIEW NORTHWARD FROM RIM OF SNOW MOUNTAIN FAULT BLOCK
Dissected plateau areas border canyon of South Fork River below South Fork Valley. River flows from right to center of view and turns northward, crossing Ochoco Range through deep notch shown in skyline.

FIGURE 2. VIEW NORTHEASTWARD ACROSS THE DISSECTED DOWNWARP OF SOUTH FORK VALLEY
River flows from right to left, beyond timbered area, across southeast limb of Mowich anticline. More sharply etched hills on left mark Triassic beds along axis of anticline. Ochoco Range forms the skyline.

JURASSIC AND TRIASSIC OF SOUTH FORK REGION

formations recognized in the main area as a whole, but there is sufficient overlapping of formations to establish their stratigraphic order without fossil evidence.

TABLE 1.—*Summary of marine Jurassic sequence of central Oregon*

Age	Stratigraphic units		Character	Thickness (Feet)
Upper Jurassic?		Lonesome formation	Black shales with intercalated beds of massive, blue-gray sandstone and conglomerate. Basal conglomerate contains pebbles of Paleozoic cherts.	4000
		—Unconformity?—		
Early Upper Jurassic		Trowbridge shale	Black shale with many large concretions. Macrocephalitid ammonites.	4000
		—Unconformity—		
Early Middle Jurassic	IZEE GROUP	Snowshoe formation	Black shale and fine-grained dark-gray laminated sandstone. Sonninid and stepheoceratid ammonites.	2800
		Hyde formation	Mainly greenish-gray and bluish-gray massive sandstone.	1080–1512
		—Unconformity?—		
	COLPITTS GROUP	Warm Springs formation	Shale and fine-grained calcareous sandstone. Sonninid ammonites.	100–300
		Weberg formation	Mainly gray granular limestone. Also calcareous sandstone, grit, and conglomerate. Rich ammonite fauna, largely sonninids and sphaeroceratids.	100–272
		—Unconformity—		
Late Lower to early Upper Lias	MOWICH GROUP	Nicely shale	Black shale with hildoceratid, amaltheid, and dactyloid ammonites.	134–228
		Suplee formation	Granular gray limestone and calcareous sandstone. Very fossiliferous.	35–150
		Robertson formation	Conglomerate, sandstone, and plicatostylid reef limestone.	150–500
		Donovan formation	Conglomerate, sandstone, shale, and limestone. Includes the fossiliferous red "Hardgrave" sandstone of other writers. May be in part equivalent to the lower Mowich group.	2241–2500

Marked differences exist between the lower and the upper parts of the Jurassic section in the main area. The lower formations, in the Mowich and Colpitts groups, are thin, calcareous, and abound in well-preserved fossils; the Hyde and later formations are thick, noncalcareous, and the

shell material is largely removed from the relatively few fossils that are present.

The Jurassic formations are essentially concordant throughout the main area, though large time intervals come between the Mowich and the Colpitts groups, and between the Snowshoe and Trowbridge formations. Erosional intervals are evident between the lower formations, for the Hyde, Weberg, and Suplee, at one place or another, bevel all earlier formations at a very low angle and lie upon Triassic or Paleozoic formations.

The sequence of formations in the main area that are known to be of Jurassic age lies with high angle discordance upon a basement of highly folded strata which are in part of Mississippian and Upper Triassic age. Some of this basement rock, however, may be Jurassic, for a sequence of shales, sandstones, and conglomerates, many thousands of feet thick, overlies the fossiliferous Upper Triassic beds but underlies the great unconformity at the base of the known Jurassic section. These are exposed between Snow Mountain and Suplee and were referred to by Schenk (1934) as "unfossiliferous strata of unknown age." When fossils are found in this series, it may be possible to determine whether the unconformity represents the Karnic to Middle Lias interval or only a part of the Lower Jurassic. These beds of uncertain age are here treated as a part of the Triassic, and references to Jurassic will apply only to the formations of known Jurassic age.

All the Jurassic formations are steeply folded, and for the most part the axes of their folds are widely divergent from the axes of the Triassic folds, so that in some places nearly vertical Triassic and Jurassic beds show angular discordance of 70 or 80 degrees. The Jurassic strata strike northeast to east over most of the area, though there are some marked and abrupt deviations from this general plan. The small faults throughout the area cause no great irregularity in the exposures except where they cut across the thin formations in the lower part of the section. The Mowich anticline, a prominent structural feature of the Jurassic formations, extends from Warm Springs Creek valley at a point 5 miles south of Suplee Post Office northeastward across the South Fork Valley at least as far as the west side of Bear Valley. It is eroded far below the base of the Jurassic section so that Triassic marine rocks and the strata of unknown age are exposed along the axis in a belt that is 2 to 6 miles wide (Pl. 3, fig. 1).

ORIGIN OF STRATIGRAPHIC NAMES

Most of the names of stratigraphic units will have little geographic significance to anyone who is unacquainted with the minor geographic features of central Oregon. An attempt to observe the rule of priority

FIGURE 1. VIEW NORTHWEST ACROSS ANTICLINAL AXIS
Morgan Mountain, a remnant of Silvies surface, rises on skyline. Triassic beds occupy most of area. Suplee and Nicely formations, on opposite limbs of the anticline, show in immediate foreground and below lava cap of Morgan Mountain.

FIGURE 2. SOUTHEAST LIMB OF ANTICLINE
Reverse of Figure 1. Shows juniper-dotted surface of Hyde sandstone in foreground, smooth hill slopes of Snowshoe and Trowbridge shales beyond, and higher points of Lonesome formation on skyline. Canyon in foreground parallels Hyde-Snowshoe contact.

MOWICH ANTICLINE IN SOUTH FORK VALLEY

FIGURE 1. PLICATOSTYLID REEF MATERIAL IN MOWICH FORMATION
Specimen below and to the right of hammer is 19 inches long and broken at both ends. Loose pieces below are typical of *Plicatostylus* specimens usually recovered.

FIGURE 2. UPPER CANYON OF SILVIES RIVER BETWEEN BEAR VALLEY AND SILVIES VALLEY
Timber on skyline marks edge of Silvies Surface. Sage-covered areas underlain by Jurassic sandstones and shales. Line of trees in middle distance marks large basalt porphyry dike.

PLICATOSTYLID REEF AND UPPER SILVIES CANYON

has eliminated some of the most appropriate names and some that might have been taken from major geographic features. Consequently, the unit names come mostly from names of ranches and minor streams.

The origins of new stratigraphic names proposed in this paper are as follows:

LONESOME FORMATION: Lonesome Creek, a tributary of Lewis Creek, in South Fork Valley.

TROWBRIDGE SHALE: H. H. Trowbridge Ranch, at the mouth of Rosebud Creek, in South Fork Valley.

Ī-ZEÉ GROUP: The Izee grange and schoolhouse near South Fork River between Poison Creek and Rosebud Creek.

SNOWSHOE FORMATION: The "Snowshoe country" on the headwaters of Silvies River.

HYDE FORMATION: John Hyde Ranch at the mouth of Poison Creek.

CŌLPITTS GROUP: Willard Colpitts Ranch, on upper Warm Springs Creek, in sec. 19, T. 18 S., R. 26 E.

WARM SPRINGS FORMATION: Warm Springs Creek.

WEBERG FORMATION: Weberg Ranch, on lower Warm Springs Creek, in sec. 18, T. 18 S., R. 26 E.

MŌWICH GROUP: Mowich Mountain.

NICELY SHALE: The old Nicely homestead on upper Beaver Creek, in sec. 29, T. 18 S., R. 26 E.

SŪ-PLEÉ FORMATION: Suplee Post Office.

ROBERTSON FORMATION: Robertson Ranch, in sec. 28, T. 18 S., R. 26 E.

DONOVAN FORMATION: Tim Donovan Ranch, in sec. 7, T. 20 S., R. 30 E.

DONOVAN FORMATION

STRATIGRAPHY

In the geologic and paleontologic literature are an unknown number of references to such things as "Hardgrave formation," "Hardgrave fauna," "Lower Jurassic fauna," and "Upper Lias fauna," in eastern Oregon. References to their localities are even more varied, and most of them indicate such vague places as "Beaver Creek, a tributary of the Crooked River," "Blue Mountains of Oregon," and "east of Red Buttes." All the references, insofar as they were applied to Oregon, appear to have had a common ancestry in a red sandstone and its fauna in the Lower Canyon of the Silvies River where the sandstone forms a minor part of an isolated section of marine rocks herein referred to as the Donovan formation. The incorrect locality references were caused largely by fossil collections passing from hand to hand and receiving incorrect locality designations. The references to "Hardgrave" result from Hyatt's (1894) correlation of the fauna with the Hardgrave fauna of Taylorsville, California.

Smith and Packard (1919) noted that Condon had obtained Lower Jurassic fossils from "Beaver Creek, a tributary of the Silvies River, and from Red Butte, north of Burns." Packard and Nelson (1921) reported that the Lower Jurassic fauna

> "was recently found to occur in a series of sedimentary and associated intrusive rocks outcropping only within the canyon of Silvies River. This series includes red impure limestone, light-colored limestone, fine-grained shale, arenaceous shale, arkosic sandstone, and basic intrusives."

The Donovan exposures (Fig. 3) occupy less than half a square mile in sections 5, 6, 7, and 8, T. 20 S., R. 30 E., 19 miles north of Burns and 10 miles west of Crow Flat Ranger Station. The area is separated from the main pre-Tertiary rock area to the north by the plateau lavas and owes its origin to the trenching of the lavas by the Silvies River and tributaries. The Jurassic sediments, and associated intrusives of basalt and basalt porphyry, formed a hill in the pre-plateau topography, and this hill was buried by rhyolite lavas and associated continental sediments during the construction of the Silvies Surface. The Silvies Canyon, over 400 feet deep, was cut across the west side of this buried hill so that Jurassic rocks were exposed on both sides of the river, and tributary gulches and canyons from the east uncovered the marine sediments for three-fourths of a mile east of the river.

Most of the Jurassic rocks occur in a single area that extends from the Donovan meadows up the spurs and gulches of the east side of the Silvies Canyon to an altitude of 400 feet above the river. Three smaller areas are isolated by rubble and intrusions near the northwest margin of the main area. The marine section is nearly surrounded, above the level of the Donovan meadows, by large intrusions, principally of the green and black basalt porphyry which is so abundant in the marine rocks elsewhere in central Oregon. Porphyry dikes and sills also cut the Jurassic section in several places. These intrusions, being a part of the buried hill, antedate the plateau lavas but nevertheless are exposed mostly between the lavas and the marine beds. The sediments strike consistently near N. 63 E. and dip northwest at angles between 40 and 65 degrees. Soil, rubble, and intrusions present the principal difficulties in working out the details of the section.

The Donovan sediments consist principally of hard, green and gray noncalcareous sandstone. Sandy shale, red sandstone, and yellowish calcareous sandstone are also well represented. Limestone and conglomerate are almost negligible. The red sandstone has been known for many years and has attracted more attention than the other deposits because of its abundant fossil Mollusca and because of its resemblance to the red Hardgrave sandstone of the Taylorsville region of California.

Red sandstones appear at several places in the area and in each place yield abundant pelecypods and a few ammonites. All the pelecypod assemblages in the various red sandstone units bear a marked but perhaps superficial resemblance to each other, particularly in the presence of species of *Pecten, Pholadomya, Pleuromya, Pinna, Modiolus,* and *Gervillia.*

The two small isolated areas on opposite sides of the river bed (Fig. 3, locs. 1, 2) appear to represent the uppermost beds of the Donovan formation. They are mostly very fossiliferous massive red sandstone with a lesser amount of gray limestone above and gray massive sandstone below. The total thickness is less than 100 feet. The fossiliferous red sandstone of lower Lime Gulch (Fig. 3, loc. 3) and its probable continuation, the small isolated outcrop of red sandstone at the head of Rattlesnake Gulch (Fig. 3, loc. 4), probably are stratigraphically below the river bed exposures. This adds about 100 feet to the section. Below the red sandstone is a lens of limestone upon which is located the abandoned lime kiln on top of the ridge. The sedimentary section, measured from the top of the limestone lens southwesterly across the main area to the lowermost beds exposed in Coral Valley, shows:

	Feet
1. Limestone lens: massive, light- to dark-gray. Contains many irregular dark streaks and spots that resemble sections of *Plicatostylus*	76
2. Sandstone: gray, green, and brown, calcareous, medium grained; weathers brown and light buff; poorly bedded. Contains terebratellid brachiopods and oyster shells	130
3. Sandstone: fine-grained, green and greenish-yellow	55
4. Concealed. Float is basalt porphyry and green sandstone	120
5. Sandstone: green, medium- and coarse-grained, massive	55
6. Pebbly sandstone: coarse-grained, well-rounded pebbles of silicified mudstone up to 2 inches in diameter	17
7. Basalt porphyry sill	40
8. Concealed. Probably green sandstone	30
9. Sandstone: green, medium-grained	32
10. Red sandstone: fine-grained, massive, very fossiliferous; outcrops in and parallel to bottom of Red Gulch	239
11. Sandstone: greenish-gray. Occasional calcareous beds are fossiliferous	64
12. Sandstone: gray and green, calcareous, fossiliferous	285
13. Sandstone: gray, massive, coarse-grained; weathers brown with tinge of red	178
14. Sandstone: similar to 13 but with a number of poorly defined pebble bands as much as 2 feet thick. Also beds of sandy conglomerate up to 6 feet thick, with well-rounded boulders of hard gray silicified mudstone. This unit and No. 13 form the backbone of the ridge between Coral Valley and Red Gulch	263
15. Mudstone and fine sandstone: flinty and silicified, mostly gray and greenish-gray, with some red; weathers to purplish-red	53
16. Basalt porphyry sill	51

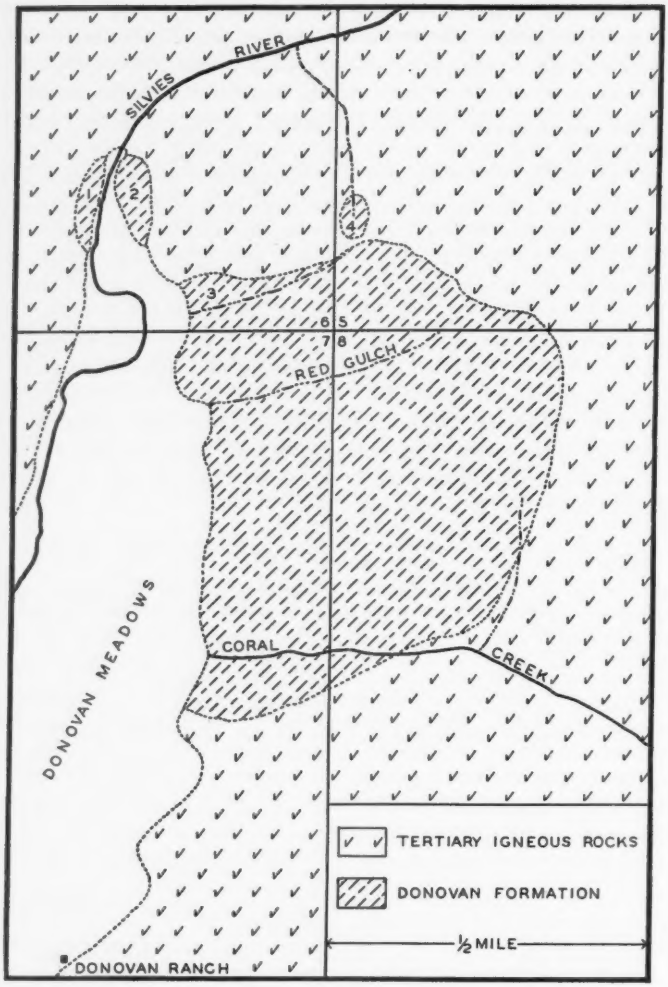

FIGURE 3.—*Distribution of Donovan formation in T. 20 S., R. 30 E.*
Locs. 1, 2, 3, 4: Fossiliferous red sandstone outcrops separated by Tertiary intrusions and lava flows.

	Feet
17. Sandstone: greenish-gray and gray, medium-grained, massive. Contains numerous granules of red mudstone and occasional waterworn pebbles and boulders of the same material up to 10 inches in diameter	37
18. Largely concealed. Shows outcrops of red mudstone and sandstone, and basalt porphyry intrusions	148
19. Partly concealed. Shows outcrops and much float of a greenish-gray sandstone which weathers tan and greenish-yellow	182
20. Grit and conglomerate: gray, calcareous matrix; waterworn pebbles of quartz up to 1 inch in diameter	13
21. Mudstone and siltstone: black and gray, fractures into splinters 1 to 3 inches long. Contains small lenses of coralline limestone. Underlies Coral Valley	362
Total	2430

Of the total thickness of 2430 feet in the measured section, 2041 feet is known to be sedimentary rock, 298 feet is concealed but probably mainly sedimentary rock, and 91 feet is Tertiary basalt sills. Thus the thickness of the sedimentary section, including the beds exposed in the isolated areas to the northwest, is at least 2241 feet and may be as much as 2500 feet.

An unconformity between beds 17 and 18 is suggested by the fact that 17 and lower beds dip at an angle of 62 degrees whereas 18 and all higher beds dip at angles less than 45 degrees. This evidence may be discounted because of the thick intrusions of Tertiary basalt, though much larger intrusives around the margins of the Jurassic area have had no appreciable effect upon the dip of the beds. Additional evidence of an unconformity is found in the abundance of red silicified mudstone and fine sandstone of bed 18 as grains, pebbles, and boulders in the sandstones and conglomerates of units 13, 14, and 17. However, present evidence does not justify recognition of two formations in the Donovan area.

AGE OF THE DONOVAN FORMATION

Fossils from the Donovan formation were first studied by Alpheus Hyatt (1894). The collection had been made by a Mr. Day, sent to Thomas Condon, then to J. S. Diller, and finally to Hyatt (McCornack, 1928). In the course of these transfers, the collection received what appears to be an incorrect locality designation, for Hyatt (1894, p. 401) stated that they came from "Beaver Creek, a tributary of the Crooked River, in the Blue mountains of eastern Oregon." Condon may have assumed that the Day collection came from the upper Crooked River because he had received previous collections of Cretaceous and Paleozoic fossils from that area. Another possibility is that the Day collection actually came from a "Beaver Creek," for Hay Creek, paralleling the lower Silvies Canyon less than 2 miles west of the Silvies Canyon area, is designated on some maps as Beaver Creek and might expose some of the red Donovan sandstone.

Washburne (1903, p. 227) attempted to correct the mistake when he stated that "These fossils did not come from Beaver Creek, but from Silver river, over thirty miles east of Beaver Creek." The conclusion that Hyatt's fauna came from the Donovan formation of the Silvies Canyon is borne out by three lines of evidence: (1) All species listed by Hyatt are found in the Donovan formation; (2) some of the original collection still in the Condon collection at the University of Oregon shows red sandstone matrix identical with the red sandstone of the Donovan; and (3) Beaver Creek, on the upper Crooked River, has been investigated more intensively than any other central Oregon pre-Tertiary area, and no one has found a similar red sandstone.

Hyatt described or mentioned the following species in the original Day collection: *Pecten acutiplicatus* Meek, *Pholadomya nevadana* Gabb, *P. multilineata* Gabb, *Pleuromya concentrica* Hyatt, *Cardinia gibbosum?* (Meek), and *Rhynchonella* sp. He concluded that the red sandstone of "Beaver Creek" was contemporaneous with the red Hardgrave sandstone of the Taylorsville region, California, which he assigned to the upper Lias. Packard and Nelson (1921) referred to the Donovan fauna as the "Hardgrave Jurassic fauna." Hertlein (1925) described a new ammonite species from the Donovan area as *"Uptonia silviesi,"* suggesting that the age of both the Donovan and Hardgrave faunas was "Charmouthian, Middle Lower Jurassic" (upper part of the Lower Lias of Arkell). He listed the following lamellibranchs from the Donovan fauna: *Anatina* sp., *Gervillia* sp., *Pecten acutiplicatus* Meek, *Pleuromya concentrica* Meek,[1] *Pleuromya depressa* Meek, *Pholadomya multilineata* Gabb, *Pholadomya* cf. *nevadana* Gabb, *Pholadomya* sp.

Little evidence can be found for the correlation of the Donovan with the Hardgrave. Of the species listed by Hyatt and Hertlein only *Pecten acutiplicatus* and *Pholadomya nevadana* have been reported in the Hardgrave fauna. At least three acutely ribbed pectens are in the Donovan fauna and none can be identified as *P. acutiplicatus.* Stages closely approximating *Pholadomya nevadana* and *P. multilineata* are present in the extremely variable series of pholadomyas in the Donovan fauna, but they have little age significance. *P. nevadana* Gabb (1869) was described from a lower Jurassic locality near Volcano in southwestern Nevada, and the work of Muller and Ferguson (1939) in this region has listed *P. nevadana* associated with two ammonite faunas which are related to Sinemurian and Pliensbachian faunas of the Lower Lias of Europe. Hyatt (1894, p. 418) reported *P. nevadana* in the Hardgrave formation which he assigned to the Upper Lias, and Crickmay (1933)

[1] *Pleuromya concentrica* Hyatt.

states that the Hardgrave is "earliest Middle Jurassic, Ludwigian" but does not list *P. nevadana*. *P. multilineata*, also in Gabb's Lower Jurassic of Volcano, Nevada, is listed by Muller and Ferguson (1939) only with a lower Pliensbachian fauna.

The principal fossils of the Donovan formation upon which the present age assignment is made are these ammonites: *"Uptonia" silviesi* Hertlein, and related species; *Coeloceras* cf. *C. pettos* (Quenstedt); *"Metechioceras"* Trueman and Williams; *Deroceras* cf. *D. impavidum* Buckman. *"Uptonia" silviesi* and related species do not meet the definition of the genus *Uptonia* Buckman but nevertheless show affinities with Polymorphitidae of the same general date as *Uptonia* and so suggest the Polymorphitan age of Buckman which corresponds to the time of the *Uptonia jamesoni* zone of Arkell and others. The *Coeloceras* also suggests Polymorphitan age, but the *Deroceras* and *"Metechioceras"* indicate the Deroceratan age, which corresponds to the time of the *Echioceras raricostatum* zone. This does not mean that both Deroceratan and Polymorphitan ages are represented in the Donovan formation, because all the ammonite genera listed above occur together in the uppermost red sandstone member of the formation and polymorphitids occur in all the red sandstone beds. Therefore, until additional information is gathered, the Donovan fauna must be assigned to a time that falls within the range of the Deroceratan and Polymorphitan ages. Strata of this date in the European sections are in the uppermost Sinemurian and lowermost Pliensbachian stages of the Lower Lias. The date proposed here applies only to the upper part of the Donovan formation in which diagnostic fossils have been found. The lower part, especially that below the possible unconformity already referred to, may be somewhat earlier. The only other known North American fauna which has been assigned to the same general date as the Donovan fauna upon satisfactory evidence is one reported by Muller and Ferguson (1939) from the upper part of the Sunrise formation in west central Nevada.

MOWICH GROUP

GENERAL FEATURES

The Mowich group includes the middle Lower Jurassic strata at the base of the Jurassic section of the main area. Though the total thickness is less than 500 feet in most places, it is divisible into three distinct formations, in ascending order as follows: the Robertson formation which normally consists of basal conglomerate overlain by greenish-gray sandstone and plicatostylid reef limestone; the Suplee formation of fossiliferous gray limestone and calcareous sandstone; and the Nicely black shale. The Robertson and Suplee may be in some part contemporaneous with

TABLE 2.—*Correlation table*

	Standard English sequence		Central Oregon	
	Stratigraphic units (Arkell, 1933)	Ages (Buckman, 1930; Spath, Neaverson, Arkell emend., 1933)	Groups	Formations
UPPER JURASSIC	Purbeck beds			
	Portland beds	Titanitan		
		Behemothan		
	Kimmeridge clay	Pavlovian		
		Pectinatitan		
		Allovirgatitan		
		Gravesian		
		Physodoceratan		
		Rasenian		
		?Prionodoceratan		Lonesome formation (Upper Jurassic or Cretaceous)
	Corallian beds	Ringsteadian		
		Perisphinctean		
		Cardioceratan		
	Oxford clay	Quenstedtoceratan		
		Kosmoceratan		
	Kellaways beds	Proplanulitan		Trowbridge shale
	Cornbrash	Macrocephalitan		
MIDDLE JURASSIC		Clydoniceratan		
	Great Oolite series	?Oxyceritan		
		Tulitan		
		Zigzagiceratan		
	Inferior Oolite	Parkinsonian		
		Stepheoceratan	Izee group	Snowshoe formation
		Sonninian		Hyde formation
	Upper Aalenian	Ludwigian	Colpitts group	Warm Springs formation
	Lower Aalenian	Canavarinan		Weberg formation
LOWER JURASSIC (LIAS)		Dumortierian		
	Yeovilian	Grammoceratan		
		Haugian		
		Hildoceratan		
	Whitbian	Harpoceratan	Mowich group	Nicely shale
	Upper Pliensbachian (= Middle Lias)	Amaltheian		Suplee formation
				Robertson formation
		Liparoceratan		
	Lower Pliensbachian	Polymorphitan		Donovan formation
		Deroceratan		
		Oxynoticeratan		
		Asteroceratan		
	Sinemurian	Microderoceratan		
		Agassiceratan		
		Coroniceratan		
		Vermiceratan		
	Hettangian	Schlotheimian		
		Psiloceratan		

the Donovan formation, but the Nicely appears to be distinctly younger. Nothing referable to the Donovan formation has been found in the main Jurassic area. Lack of deposition, erosion, or marked lateral changes of lithology and thickness may well account for the apparent absence, but it is also possible that the Donovan lies in the folded basement rocks below the base of the Mowich group.

Mowich exposures are found principally in the "Suplee-Izee area" in the western half of the pre-Tertiary area where they trend northeast between Suplee Post Office in the upper Crooked River country and Izee Post Office in the South Fork valley. Far to the northeast is the much smaller "Seneca area" of Mowich exposures, 4 miles north of Seneca on the north side of Bear Valley.

The Mowich exposures in the Suplee-Izee area are distributed in two long narrow bands, 2 to 6 miles apart, along opposite limbs of the truncated Mowich anticline. The exposures are not continuous because both the Colpitts and Hyde formations extend across the Mowich in places and lie upon the Triassic formations. On the south limb all three formations of the Mowich come in between the Triassic and Colpitts where the Suplee-Izee road crosses the Warm Springs Creek-Beaver Creek divide 7 miles south of Suplee. They extend eastward to Mowich Mountain where they pass beneath a northward-projecting promontory of plateau lavas. The Mowich anticline reappears from beneath the lava cap on the north side of Snow Mountain 4 miles east of Mowich Mountain, but there the Mowich group is absent and the Middle Jurassic Hyde formation lies upon the Triassic beds. About 4 miles northeast of Snow Mountain, on the east side of Big Flat, the Suplee and Nicely come in between the Hyde and the Triassic and continue in that relation northeastward into and across the South Fork valley as far as Poison Creek where they again pinch out beneath the Hyde. On the north limb only the Robertson is exposed from Beaver Creek canyon northeastward to the South Fork, but northeast of the South Fork valley the Suplee and Nicely appear on Morgan Mountain.

The Mowich group is steeply folded so that in most localities it is inclined more than 45 degrees and in some places is nearly vertical. The Mowich exposures on both limbs of the anticline show many irregularities, due to faulting and local warping, which can be shown only by detailed mapping. None of these irregularities is of sufficient magnitude to deflect the outcrops more than half a mile from the course of the anticlinal limbs indicated in Figure 1.

Between the eroded limbs of the Mowich anticline a very thick series of nearly vertical Upper Triassic shales with resistant bands of sandstone, conglomerate, chert, and limestone strike northward and show

high-angle discordance of both dip and strike with the overlying Mowich group. In the western part of the Suplee-South Fork area the resistant Triassic beds, separated by thick shale beds, form a grid of ridges transverse to those of the resistant lower Mowich group on opposite limbs of the anticline. Sills of basalt porphyry accentuate the topographic expression of the Triassic and Jurassic structures. In the Seneca area, likewise, the Mowich truncates Triassic marine sediments with high-angle discordance.

The subdivision of the Mowich into three formations does not rest only upon the marked differences in lithology and geographic distribution mentioned. All three formations are very fossiliferous, but no species common to two formations has yet been found. The differences between the various parts of the Mowich may be the result of superposition of distinct but contemporaneous lithologic and faunal facies by an advancing sea, but the differences in geographic distribution of the Robertson with reference to the other two formations has cast doubt upon its inclusion with the same transgressive phase.

ROBERTSON FORMATION

The Robertson includes, in most places, two distinct lithologic facies: (1) a basal section of conglomerate and coarse greenish-gray sandstone upon the truncated beds of the Upper Triassic, and (2) an overlying reef limestone. In some places, however, the differentiation is not complete so that lenses of reef limestone are found in sandstone, and lenses of sandstone are found in the limestone. The thickness of the Robertson is about 250 feet in most places, and the maximum observed is about 500 feet.

Apparently the thickness and composition of the coarse clastic beds of the Robertson are related to the distribution of coarse clastics in the underlying Triassic formations. The Triassic-Robertson unconformity bevels many conglomerate and coarse sandstone beds along the southwestern part of the Mowich anticline, and there the basal Robertson is thick and conglomeratic. Northeastward, on Pine Creek, the Triassic section contains fewer conglomerate beds, and here coarse basal beds are absent. The coarse materials in the Triassic clastics are mostly chert with minor amounts of limestone derived largely from older Triassic and Paleozoic formations like those exposed south and west of Suplee (Packard, 1928). The composition of the Robertson conglomerates and sandstones is similar to that of the Triassic clastics, and, though they also may have been derived in large part directly from the Paleozoic formations, it seems more probable that they were reworked from poorly consolidated Triassic sediments. Grains and pebbles of these Paleozoic

cherts also appear in successively later Jurassic formations as well as in the Tertiary sediments.

The reef limestone overlying the basal member is composed largely of a gregarious rudistidlike pelecypod, *Plicatostylus gregarius* Lupher and Packard, which flourished on the seaward side of the inshore zone of coarse basal deposits. Though much of the reef phase is a limestone with many elongate plicatostylid valves, there are many variations from this type. The shell limestone is composed either of closely packed individuals that grew in parallel position or of a tumbled mass of broken shells. Both types grade through decreasing amounts of shell material into either fine gray limestone or greenish sandstone. Gradations from fine limestone to sandstone also are common, so that almost any combination of these extremes—shell limestone, fine limestone, and sandstone—may be found in the Robertson formation. A large gastropod of the genus *Nerinea* also contributes materially to the limestone in a few places.

SUPLEE FORMATION

This formation is composed of gradations from calcareous gray medium- and fine-grained sandstone to gray granular limestone. It is mostly poorly stratified or massive, and cut by numerous joints. In only a few places do the exposures stand out in rocky ledges; for the most part they appear on smooth hillsides where they are recognizable from a distance by their buff and light-brown weathered surfaces. The thickness ranges from 35 feet to 150 feet.

The Suplee lies upon the reef limestone of the Robertson in the type area and Bear Valley area, but elsewhere the Robertson is absent so that the Suplee lies discordantly upon Triassic beds.

Fossils, mostly pelecypods, are abundant in the Suplee. A large, acutely ribbed pecten, similar to *P. acutiplicatus* Meek and to the large pectens of the Donovan formation, is an abundant, widespread, and reliable marker for the Suplee formation. Nothing similar to it has ever been found in other formations of the main Jurassic area.

NICELY SHALE

Lithologically the Nicely is fairly uniform both vertically and laterally, being dark-gray to black sandy shale. Occasional spheroidal black calcareous concretions, some as much as 2 feet in diameter, have seldom failed to produce well-preserved ammonites and other fossils. Measurements in areas other than those in which it pinches out beneath younger formations give a range in thickness from 134 to 228 feet. The Nicely lies conformably upon the Suplee formation throughout its known occurrence. It is overlain by the Colpitts group in the type area and by the Hyde

formation eastward in the South Fork Valley. It is less resistant to erosion than the underlying and overlying formations, and consequently occupies, in most places, a soil-covered trough.

MOWICH GROUP OF THE TYPE AREA

General features.—The type area of the Mowich group, and its three component formations, is located along the headwaters of the South Fork of Beaver Creek, 7 miles southeast of Suplee Post Office, in sections 26, 27, 28, and 29, T. 18 S., R. 26 E. The narrow band of exposures appears from beneath the Tertiary lavas on the west side of Mowich Mountain and extends westward 3½ miles, with some interruptions and repetition by faults, as far as the Beaver Creek-Warm Springs Creek divide. This is the west end of the deeply eroded Mowich anticline on which the Mowich group dips southward mostly at angles of 50 to 65 degrees. Underneath are the nearly vertical Triassic marine beds with northward strike widely divergent from that of the Mowich. The hard conglomerates and sandstones in the predominantly shale section of the Triassic produce conspicuous northward-trending ridges which end abruptly on the south against the eastward-trending ridge formed by the lower part of the Mowich group. The upper Mowich (Nicely) is overlain unconformably by the early Middle Jurassic Colpitts group. The two formations appear to be parallel where they outcrop together, but on the Warm Springs Creek-Beaver Creek divide, in section 29, the Colpitts extends across all the Mowich formations in less than half a mile and lies upon the steeply inclined Triassic beds.

The line of Mowich outcrops near the east end of the type area is offset about half a mile to the south by a cross fault at the west base of Mowich Mountain. The west end of the outcrops is also offset to the south by another fault which lies about one-fourth mile east of the old Nicely ranch house. Immediately east of this latter fault the lower part of the Mowich is repeated in several small fault blocks north of the main outcrops which lie along the limb of the anticline.

The following section, measured near the Robertson ranch house in section 28, will illustrate the general character of the group.

	Feet
Nicely shale	134
Dark-gray and black sandy shale. Contains large fossiliferous concretions with *Amaltheus, Platyharpites, hildoceratids,* and *dactyloids.*	
Suplee formation	40
Fine-granular gray limestone and calcareous sandstone weathering light brown and buff. Contains large *Pecten,* rhynchonellid, and other typical Suplee species.	
Robertson formation	197
Coarse sandstone and fine conglomerate with subangular to rounded pebbles of chert and limestone in a matrix of very calcareous sandstone	24

	Feet
Dark gray and greenish-gray limestone grading into sandy limestone. Weathers buff. Contains *Nerinea* and *Modiolus*	28
Hard, white reef limestone with numerous plicatostylid shells	10
Basal conglomerate. Fine, poorly sorted conglomerate with angular pebbles of chert and limestone; matrix of coarse sandstone; color dark gray to brown	135
Total	371

Robertson formation.—Basal conglomerate and sandstone is well developed in the type Robertson and, except in the measured section already given, it is fairly well differentiated from the upper limestone phase. The thickness of the basal beds in the type area is less than 150 feet, except on the west side of Mowich Mountain where it appears to exceed this figure considerably. These beds are more resistant to erosion than the other members of the formation and form the high part of a strike ridge that is especially conspicuous where the Mowich lies upon thick, easily eroded Triassic shale beds. The most common type of sediment in the basal phase is a greenish-brown, poorly sorted mixture of coarse sand and pebbles with occasional cobbles as much as 6 inches in diameter. The grains and pebbles are colored cherts with minor amounts of limestone and porphyry. The porphyry is very abundant in certain localities, as on the west side of Mowich Mountain where the basal beds contain porphyry cobbles as much as 10 inches in diameter.

The limy phase of the Robertson is characterized by various proportions of reef limestone and sandstone. The limestone is made up largely of *Plicatostylus gregarius* Lupher and Packard. This species was a sessile and gregarious pelecypod whose elongate lower valve resembled, in proportion, the most elongate rudistids of the Cretaceous. They were attached to one another or to the substratum and grew upright to lengths of more than 20 inches. In places they form reefs composed of closely packed parallel columnar valves. The small crevices between valves are filled with limestone or sandstone and fragments of broken shell. In other places the valves are so distantly spaced that the filling between valves constitutes as much as 75 per cent of the rock, or the plicatostylids may be confined to lenses in sandstone. Most of the shells are still upright, and some dip surfaces show innumerable oval cross sections of this peculiar fossil. A large triplicate *Nerinea* is abundant locally, mostly in limestones that contain few plicatostylids. The matrix is so hard that the fossils can be broken away in only a few localities. The thickness of the reef member is variable in the type area; the maximum is about 100 feet—a moderate thickness in comparison to some other localities.

The plicatostylids apparently flourished on the firm substratum of sands and gravels of the basal beds. Lenses of coarse sandstone and fine

conglomerate in the reef material indicate that the shells grew where the wave action was strong. The reef limestone forms a continuous, well-defined member over all but the westernmost end of the type area. It resembles a biostrome, but the fairly continuous layer probably resulted from the advance of relatively restricted reef-building areas during the transgression of the Mowich sea.

Suplee and Nicely formations.—The characters of the Suplee and Nicely of the type area conform closely to the general description already given so that little needs to be added. The thickness of the type Suplee, 40 to 60 feet, is less than in other sections, but the formation is much more fossiliferous than elsewhere. Rhynchonellid brachiopods are profuse but show little variety. Pelecypods are abundant and show the greatest number of species. Next in abundance are corals, cephalopods, Bryozoa, gastropods, and crinoids.

MOWICH GROUP OUTSIDE THE TYPE AREA

South Fork Bridge.—The continuation of the Mowich group northeast of the type area along the south limb of the Mowich anticline is obscured by Tertiary lavas and by the Middle Jurassic Hyde formation except in the South Fork Valley between Big Flat and Poison Creek. The Robertson formation is absent here, and the Suplee lies upon the truncated Triassic beds. A section measured on the high hilltop immediately west of South Fork Bridge gave the following Mowich section:

	Feet
Nicely shale with concretionary layer at top and typical fauna	228
Suplee formation of green and gray calcareous sandstone and gray limestone; brown and buff on weathered outcrop; typical fauna	134
Robertson formation	absent
Total	362

About a mile northeast of the above section, on the east side of Poison Creek Canyon, the Suplee and Nicely members pinch out, and the younger Hyde formation lies directly upon the Triassic sediments. The Mowich has not been seen farther to the northeast in the South Fork region or in the headwaters of the Silvies River on the west side of Bear Valley.

Swamp Creek.—The Mowich exposures on the north limb of the Mowich anticline show exceptional developments of the Robertson formation between Beaver Creek Canyon and the South Fork River. Neither the Suplee nor Nicely formations have been found here. Their absence is probably caused by the pre-Colpitts erosion so well shown in the type area.

Good exposures of the Robertson formation are present where upper Swamp Creek and its tributaries cut through the limb of the anticline.

The best locality in this area is in a small gully, about half a mile southwest of Swamp Creek, in section 3, T. 18 S., R. 26 E. The total thickness of the Robertson there is 246 feet; of this 133 feet is basal conglomerate and 113 feet is limestone. The basal conglomerate is composed of rounded pebbles mostly less than an inch in diameter in a matrix of grit and coarse sandstone. Near the base are a number of limestone and chert boulders which apparently were derived directly from the underlying Triassic conglomerates. The reef limestone of the Robertson formation is composed almost entirely of *Plicatostylus* shells packed solidly together in parallel positions (Pl. 4, fig. 1). This is the best *Plicatostylus* locality known; the elongate lower valves break apart readily, and many pieces, 10 to 20 inches long, can be extracted.

Pine Creek.—The exceptional development of the Robertson formation continues northeastward along the north limb of the anticline from Swamp Creek into the drainage basin of Pine Creek, a tributary of the South Fork of the John Day. South of the Harris ranch house, in section 26, T. 17 S., R. 26 E., the Robertson section shows about 400 feet of basal conglomerate and more than 100 feet of reef limestone. Two and a half miles northeast of the Harris locality the Robertson is well exposed for 2 miles along the north side of the Pine Creek Canyon as far as the mouth of Brisbois Creek. The Colpitts group, which is present above the Robertson at the Harris ranch, is apparently absent in the Pine Creek Canyon so that the Robertson is overlain by Hyde shales and sandstones.

The Pine Creek Canyon exposures differ markedly from the other Robertson sections. Basal conglomerate is absent so that the section begins with sandstone and limestone. This unusual condition probably resulted from the small amount of coarse clastic material in the underlying Triassic beds. Other differences are the alternation of sandstone and limestone beds, and the predominance of dense, massive limestone over shell limestone. The following measured section, taken at the junction of Spring Creek and Pine Creek in section 18, T. 17 S., R. 27 E., will serve to illustrate the unusual features of these Robertson exposures:

	Feet
Sandstone, massive, coarse- and medium-grained, greenish-gray, calcareous	24
Limestone, massive, dark gray, siliceous; contains a few plicatostylid shells	4
Sandstone, massive, medium-grained, calcareous, light greenish-gray	18
Limestone, dark gray, siliceous	6
Sandstone, massive, greenish-gray, mottled, calcareous; interbedded with some gray calcareous sandstone	41
Gray and green calcareous sandstone alternating with beds of gray limestone	31

	Feet
Shell limestone. A prominent white ledge with closely packed, broken plicatostylid shells	23
Sandstone, gray, medium-grained, finely mottled	19
Limestone, dense, gray, siliceous, in beds ranging from 1 to 2 feet in thickness	20
Sandstone, greenish-gray, medium-grained, massive, calcareous	23
Greenish-gray, medium-grained, calcareous sandstone and sandy limestone	4
Limestone, dark gray and black, dense siliceous	3
Sandstone, coarse- and medium-grained, greenish-gray	18
Mostly flinty dark gray siliceous poorly bedded limestone. Also calcareous greenish sandstone in thin beds. A limestone bed 10 feet above the base yielded a large Robertson *Nerinea*	48
Total	282

Morgan Mountain.—An important section of the Mowich group is present on Morgan Mountain 3½ miles northeast of the lower Pine Creek area. Morgan Mountain is a lava-capped remnant of the Silvies Surface which juts out from the north side of the South Fork canyon (Pl. 3, fig. 1). This section is unlike other north limb sections in that the Robertson formation is absent and the Suplee and Nicely are present. Excellent exposures trend northeast across the south face of the mountain immediately below the basalt lava rim. The Suplee lies unconformably upon Triassic strata. It is composed of fossiliferous brown and greenish calcareous sandstones and has a thickness of about 150 feet. The Nicely shale differs from that of the type area largely in the greater number of spheroidal concretions and in the presence of thin shell-limestone beds composed of ammonites and brachiopods. Its thickness, approximately 200 feet, is about the same as shown in other areas.

Seneca area.—On the north side of Bear Valley, 4 miles north of Seneca, are three small areas of Mowich exposures totaling less than a square mile. All three formations are present here but show some marked differences from the exposures of the Suplee-Izee area. The beds dip mostly at low angles and lie as remnants upon the Upper Triassic shales and sandstones. High-angle discordance between Upper Triassic and Lower Jurassic is present here as elsewhere in central Oregon.

The western part of the largest Mowich area, in section 10, T. 16 S., R. 31 E., shows a gentle, northeastward-trending anticline of Robertson formation. The basal part is black shale, totaling about 60 feet in thickness. The basal shale is succeeded by an exceptional thickness of light-gray limestone in excess of 200 feet. The usual plicatostylids and nerineas are abundant but are a minor part of a predominantly limestone section. Farther east, in section 11, the Robertson limestones have been greatly disturbed so that nearly vertical dips and complicated struc-

tures prevail. A brown calcareous sandstone with the Suplee fauna was found in one place apparently stratigraphically above the Robertson limestone. Black shale similar to the Nicely shale is also present and has yielded echioceratid ammonites in association with giant belemnites and ichthyosaur vertebrae, but its position in the section could not be determined. It is possibly equivalent to the basal shales.

The Mowich strata in the two small areas half a mile northwest of the main area are small remnants preserved in down-dropped blocks bounded on the east and northeast by faults toward which the beds dip at angles of less than 30 degrees. The basal part of the section is concealed, although some of the Robertson limestone is exposed, and above this comes the normal sequence of Suplee and Nicely formations. The Suplee, mostly limestone and calcareous sandstone, contains the large acutely ribbed *Pecten*, small rhynchonellid brachiopods, and other typical Suplee species. The Nicely appears to be represented by a black shale overlying the Suplee but no fossils were found to confirm this identification.

AGE OF THE MOWICH GROUP

The age assignments here proposed for the three formations of the Mowich group can be treated best collectively. The Robertson formation is set off from the Suplee and Nicely formations by marked differences in distribution already noted. The absence of the Robertson in the South Fork Bridge area and on Morgan Mountain does not necessarily mean that the Robertson was eroded before the deposition of the Suplee; its absence may be accounted for by local failure of environmental conditions conducive to the development of basal Mowich clastics and associated reef limestones. The only identifiable Robertson fossils are the extremely abundant *Plicatostylus* and *Nerinea*, which have little age significance. *Plicatostylus gregarius* Lupher and Packard is a highly aberrant pelecypod of unknown affinities. Its only certainly known occurrence is in the Robertson formation of central Oregon. Fragmentary specimens from the limestones of the Donovan formation may represent this species, and a very similar and possibly identical species is reported by Muller and Ferguson (1939, p. 1621) in the Dunlap formation of western Nevada. The age of the Nevada species has not been determined more closely than late Lower Lias or Middle Lias. The *Nerinea* is a new species, similar in some respects to *N. thompsonensis* Crickmay, of probable lower Middle Jurassic age, from the Thompson limestone of Taylorsville, California. Similar forms are found in the *N. oolitica-parva-attenuata* group of the lower Middle Jurassic of England. The evidence of this species of a characteristic Middle Jurassic genus might have been given some weight if the Robertson formation

had not been found beneath the early Upper Lias faunas of the Suplee and Nicely formations.

Fragmentary remains of what appear to be *Plicatostylus gregarius* in the Donovan formation suggest that the Robertson is nearer in age to the Donovan formation than to the other Mowich formations. The echioceratid ammonites present in shales associated with the Robertson limestones of the Bear Valley area strengthen this idea because the Echioceratidae have about the same chronologic position, in other parts of the world, as relatives of the Donovan ammonites. Marked differences in lithology and thickness of the Robertson and Donovan formations, though separated by a distance of only 21 miles, oppose contemporaneity. Therefore, the only conclusion regarding the age of the Robertson that can be made at present is that it is older than the overlying Suplee formation, and probably not older than the Donovan formation. The time range within which it probably comes is Deroceratan to Amaltheian ages inclusive, a time that corresponds to the uppermost Sinemurian and the Pliensbachian stages of Europe.

The Suplee and Nicely formations apparently have no species in common, but ammonites of both formations indicate the same general date. The Suplee ammonites lack some details necessary for exact identification but they are species of Dactylioceratidae and early Hildoceratidae indicative of Harpoceratan age. Some of the Suplee pelecypod species show some affinity to the Donovan fauna. The large *Pecten*, a common and widespread marker for the Suplee formation, is apparently the same as the one in the red sandstone of the Donovan which Hyatt (1894, p. 418) referred to as *"Pecten acutiplicatus* Meek." Several *Pholadomya* specimens from the Suplee can be nearly duplicated in the variable series of pholadomyas in the Donovan formation. Little reliance is placed upon these pelecypods, because similar forms are distributed rather generally through the Lower Jurassic elsewhere.

The abundant and well-preserved ammonites of the Nicely shale give more definite results than other Mowich faunas and fix an upper time limit for the Robertson and Suplee formations. Among the Nicely shale ammonites are these significant forms:

Several species of early Dactylioceratidae
Amaltheidae
Amaltheus, n. sp., cf. *A. reticularis* (Simpson)
Amauroceras, n. sp.
Early species of Hildoceratidae including:
Platyharpites, n. sp.
Harpoceras, several new species.

The *Harpoceras* species and early Dactylioceratidae agree with the Suplee ammonites in indicating Harpoceratan age, but *Amaltheus, Amau-*

roceras, and *Platyharpites* indicate the later part of the preceding Amaltheian age. That these do not represent stratigraphically distinct faunas of Amaltheian and Harpoceratan ages is shown by the occurrence, in the same concretion, of *Amauroceras*, *Platyharpites*, and *Harpoceras* stratigraphically above all other species, and the association of *Amaltheus* with early Hildoceratidae and Dactylioceratidae near the base of the formation.

Thus the Suplee and Nicely ammonites indicate a somewhat later date than the Donovan fauna—a date that would probably come within the time range of the late Amaltheian and the Harpoceratan ages or some unknown intermediate time. In terms of Arkell's nomenclature this is the time of the *Paltopleuroceras spinatum*, *Dactylioceras tenuicostatum*, and *Harpoceras falcifer* zones, ranging from late Middle Lias into the early Upper Lias.

COLPITTS GROUP

GENERAL FEATURES

The Colpitts group is an early Middle Jurassic sequence of limestone, sandstone, and shale which crops out only between Snow Mountain and Suplee in the western end of the pre-Tertiary rock area. The formation is less than 500 feet thick in most places but it is the most fossiliferous formation in the central Oregon Jurassic section and thus is of considerable paleontologic importance. Along the western part of the Mowich anticline, the Colpitts group lies upon the Mowich group and below the Hyde formation without appreciable angular discordance above or below. The Mowich group is absent along Warm Springs Creek and northward in the Camp Creek valley as far as Suplee, and the Colpitts lies upon Triassic and Paleozoic beds with high-angle discordance. Field evidence indicates that the Mowich was removed by erosion before the deposition of the Colpitts, and this is to be expected because the faunas show a large hiatus between the two groups. No such large hiatus separates the Colpitts from the overlying Hyde formation, but, as previously noted, the Hyde extends across both Colpitts and Mowich in the South Fork area and lies upon Triassic beds. Sharp flexures and small faults are common along the Colpitts exposures.

The Colpitts group is divided into two formations: the lower or Weberg formation of limestone and calcareous sandstone, and the upper or Warm Springs formation of fine sandstone, shale, and mudstone. The group may exceed 800 feet in thickness on Freeman Creek at the southeastern side of the Colpitts area. Both formations thin toward the northwest. The total thickness on Warm Springs Creek is about 400 feet and in the vicin-

ity of Suplee at the northwest side of the Colpitts area it is less than 200 feet.

WEBERG FORMATION

This formation is largely limestone with various admixtures of sand. With increasing sand content the limestones grade into calcareous sandstones. Both types are gray on the fresh surfaces; in general, the fairly pure limestones are very light gray, and darker shades of gray become more pronounced with increasing sand content. Light browns and buffs prevail on weathered surfaces; no Weberg limestone shows white outcrops like those of the Robertson formation. The Weberg beds are sufficiently resistant to remain above the alluvium, and they crop out mostly on smooth hillsides which show bedrock at the surface or beneath only a thin soil cover. The Weberg is almost entirely limestone and medium- and fine-grained sandstone where it lies upon the Nicely shale along the Mowich anticline. The formation contains much reworked chert on Warm Springs Creek and northward into Camp Creek where it lies above coarse clastics and cherts of the Paleozoic and Triassic formations. The coarse material, ranging from coarse sand to small pebbles, is largely concentrated near the base of the formation, but much of it is present throughout the section. Coarse basal beds are especially noticeable in the thin northern part of the Weberg section on Camp Creek, where a true basal conglomerate, about 20 feet thick, crops out along the east side of the creek. These features illustrate the control of the Weberg lithology by the lithology of beds below the unconformity. The northwestward thinning of the Weberg is shown by a thickness of 225 feet on Warm Springs Creek, and about 100 feet on Camp Creek.

The Weberg contains an extraordinary abundance and variety of well-preserved marine fossils. Pelecypods and brachiopods are most abundant numerically and in places are so closely packed that the beds approach shell limestone. The ammonites, while less numerous, show a great variety of species. Characteristic of Weberg exposures are loose fragments of large, coarsely ribbed Sonninidae. Other invertebrates, minor in number, include large nautiloids, belemnoids, and gastropods. Reptile bones also are present in several localities associated with marine fossils in sandy limestones. The fossils show a definite stratigraphic arrangement which will permit recognition of several zones, but, as the Colpitts fauna as a whole is of very different aspect from others previously reported in North America and contains largely new forms, the treatment of these zones must await completion of paleontologic studies now in progress.

WARM SPRINGS FORMATION

This formation represents a fine sandstone, shale, and mudstone gradation in the upper part of the Colpitts. The contact between the Warm Springs and Weberg formations is not sharply defined, but the Warm Springs can be recognized by the absence of limestone, except for occasional lenses less than a foot thick, by the abrupt decrease in the grain size of the sandstones and the appearance of sandy shale and mudstone, and by the change of outcrops from the buff of the Weberg to the more somber gray of the Warm Springs.

The Warm Springs formation is easily eroded. Though the lower and middle parts are well exposed on the lower slopes of hills and ridges formed by the more resistant Colpitts group, the upper part and the shales at the base of the overlying Hyde formation are confined to soil-covered depressions. Consequently, the exact thicknesses of the Warm Springs in various areas cannot be determined, although a northwestward thinning of the formation is evident. The thickness on Freeman Creek is not less than 300 feet; on Warms Springs Creek it is about 200 feet; and near Suplee in the Camp Creek area it is about 100 feet.

The fossils of the Warm Springs formation are principally ammonites. Most of these are very different from the Weberg ammonites, but occasional thin lenses of limestone have yielded Sonninidae of the Weberg type, suggesting that the faunal differences are largely facies differences and that temporary recurrence of lime facies in the Warm Springs permitted the invasion of the large Weberg Sonninidae. Also present are a few pelecypods, belemnites, and reptile bones.

COLPITTS GROUP OF THE TYPE AREA

The type area of the Colpitts group is on the east side of Warm Springs Creek valley in sections 19, 20, 29, and 30, T. 18 S., R. 26 E. Here the formation lies across the narrow southwest part of the eastward-trending Mowich anticline. Near the south side of the type area, both the Mowich and Colpitts groups on the south limb of the Mowich anticline swing around to a northwest strike in a faulted area immediately south of the point where the Suplee-Izee road crosses the Warm Springs Creek-Beaver Creek divide in section 29. All formations of the Mowich group pinch out within half a mile northwest of that point, and the Colpitts lies upon the steeply inclined Triassic beds. In that relationship it continues northwest and north across the Mowich anticline as far as Suplee Hot Springs, where it dips steeply northward beneath the Hyde formation on the north limb of the anticline. Minor folds are present, but in general the Colpitts dips southwestward beneath Warm Springs Creek at angles

between 5 and 40 degrees and underlies most of the east side of the valley. Gullies in this thin formation have exposed several windows of Triassic sediments; two large windows are present in section 19 where Warm Springs Creek has cut across two eastward trending minor anticlines. The Weberg formation, being more resistant than the Warm Springs formation, is extensively exposed from the Warm Springs-Beaver Creek divide westward to Warm Springs Creek, whereas the overlying Warm Springs formation is largely stripped off the valley slopes and concealed beneath soil and alluvium.

The Weberg formation is mostly light-gray limestone with varying amounts of arenaceous material, and a lesser amount of calcareous sandstone. Its maximum thickness is about 225 feet and the minimum thickness, where the upper beds are uneroded, is about 175 feet. A thin basal conglomerate is present near the Colpitts ranch but elsewhere in the type area the formation begins with limestone or coarse calcareous sandstone which contains many coarse grains and granules of chert and fragments of shells. The Warm Springs formation is composed of dark-gray fine-grained sandstone, shale, and mudstone. Its exposed thickness is less than that of the Weberg formation, but the total may be somewhat greater. The best Warm Springs exposures are on the south side of the type area along the Suplee-Izee road which follows this formation from the Colpitts ranch house to the Beaver Creek divide.

A complete section of the Weberg formation was measured in a gully about one-fourth mile northeast of the Colpitts ranch house in the NW ¼ of section 30. The Warm Springs formation is present here but was not measured because of poor exposures. The Weberg sequence, in descending order, is:

	Feet
Dark-gray calcareous fine sandstone	19
Light-gray limestone with several fossiliferous bands containing sonninid, sphaeroceratid, and strigoceratid ammonites, the ammonite genera *Zugophorites* and *Zugella* of unknown family relationships, and pelecypods	65
Gray limestone which weathers brownish yellow. Contains large terebratellid brachiopods and pelecypods	13
Very light-gray limestone with granules of chert	8
Gray sandy limestone which weathers light buff and brown. Contains pelecypods and numerous rhynchonellid brachiopods	27
Brown limestone with several beds of very calcareous coarse sandstone. Contains abundant pelecypods and sphaeroceratid ammonites	22
Hard, gray and brown sandy limestone with numerous granules of chert	19
Light-gray limestone with numerous shell fragments and granules of chert. Lies upon nearly vertical "Triassic" conglomerate and coarse sandstone	17
Total	190

COLPITTS GROUP OUTSIDE THE TYPE AREA

South limb of Mowich anticline.—Colpitts exposures extend eastward from the type area as two narrow bands, one along each limb of the Mowich anticline. Those on the south limb are offset and repeated by faults, but they trend generally eastward from the south side of the type area to Mowich Mountain and lie mostly in the southern parts of sections 29, 28, 27, and 26 between the Ochoco National Forest boundary and the Suplee-Izee road. The exposures are poor in sections 29 and 28 because of proximity to the edge of forest-covered lava plateau, but there are several outcrops of the lower part of the Weberg formation with abundant fossils. Farther east, a prominent ridge, extending south from the Robertson ranch house in section 28 exposes the Weberg and part of the Warm Springs formations. This Robertson section, and its continuation eastward across Freeman Creek to Mowich Mountain, presents an unsolved problem that may have some bearing on the Mowich group, because beds that cannot be referred with certainty to either the Mowich or Colpitts groups separate the upper Mowich (Nicely shale) and the typical Weberg formation of the Colpitts group. Elsewhere the lowermost Weberg is marked by a very characteristic fauna, but in the Robertson and Freeman Creek area this fauna occurs 260 feet above the Nicely shale. The intervening beds, comprising a thickness greater than the entire Weberg section of the type area, are composed of gray and greenish-gray unfossiliferous sandstones. Lithologically they resemble the Weberg more closely than the Mowich but pinch out within half a mile west of the Robertson section. They might well be post-Nicely beds of the Mowich coming in beneath the Colpitts-Mowich unconformity, because that unconformity bevels all the Mowich west of the Robertson area. For the present, these beds of doubtful relationships may well receive the designation of "incertae sedis."

The Colpitts group underlies most of the Freeman Creek basin west of Mowich Mountain, but the bedrock is so obscured by soil and alluvium and so broken by faults that little important information has been gained here. The Weberg formation is repeated twice, probably by faults, on the south side of the basin within 2 miles of the Mowich anticline exposures. The Warm Springs formation is largely concealed, though road cuts along Freeman Creek furnish exposures that indicate a thickness of over 300 feet.

North limb of Mowich anticline.—The Colpitts outcrops on the north limb extend from the north side of the type area on Warm Springs Creek northeastward for over 6 miles to upper Pine Creek. The Hyde formation overlies the Colpitts all the way. Triassic beds underlie the forma-

tion on Warm Springs Creek, but in the Beaver Creek canyon, and northeastward to Pine Creek, the Robertson formation of the Mowich group lies between the Colpitts and Triassic.

The Colpitts beds of this area are lithologically similar to those of the type area. The gray limestones of the Weberg formation are exposed on smooth hillsides, and the fine sandstones and shale of the Warm Springs formation are largely covered by soil and alluvium. A measured section of the Weberg on a western tributary of Swamp Creek, in section 3, T. 18 S., R. 26 E., is as follows:

	Feet
Fine-crystalline gray limestone weathering light buff to gray	90
Gray limestone with small sonninids	2
Dark-gray impure limestone and fine-grained calcareous sandstone; ammonites abundant	16
Gray limestone. Contains sonninids and other ammonites	7
Fine crystalline gray limestone grading down into gritty limestone	62
Gray limestone with chert granules and brownish and dark gray sandy limestone	14
Dark-gray limestone weathering brown and brownish-buff. Contains pelecypods	24
Very light-gray limestone containing many shell fragments	43
Basal bed of fine-grained sandstone lying upon the Robertson formation of the Mowich group	14
Total	272

The Colpitts exposures at the Harris ranch house on upper Pine Creek, section 26, T. 17 S., R. 26 E., show an exceptional development of the lowermost zones of the Weberg formation. The total thickness of the formation is about 175 feet and it is composed of characteristic light-gray sandy limestone and a minor amount of gritty limestone. The abundant fossils show the characteristic fauna of the type area.

Camp Creek.—The Camp Creek area includes the north limb of a syncline that adjoins the Mowich anticline on the north, a northward extension from the syncline along the east side of Camp Creek, and various remnants on the west side of the valley near Suplee Post Office. The exposures north of the syncline show an abrupt change in the axes of folding, from N. 60 E. on the syncline to north and northwest. Both formations are present here but they are much thinner than in the type area, each formation being, in most places, less than 100 feet thick. The Weberg formation includes limestones and calcareous sandstones similar to those in the type section but contains more coarse grains and granules of chert. In most places the basal beds are hard, coarse, and low in lime content because of the abundance of chert derived from the Paleozoic and Triassic formations. A 20-foot basal conglomerate crops out on the east side of Camp Creek. It contains pebbles and cobbles of limestone and chert derived from Triassic conglomerate beneath.

AGE OF THE COLPITTS GROUP

The abundant ammonites of the Colpitts group clearly indicate a date within a time that corresponds to the Upper Aalenian and Lower Bajocian stages of Europe. This time may be regarded either as earliest Middle Jurassic, or as latest Lower Jurassic and earliest Middle Jurassic.

The Sonninidae, which are very abundant in the Colpitts group, are characteristic of Lower Bajocian faunas in other parts of the world, especially in England where they are a conspicuous element in the faunas of "Sonninian age." Other ammonite families, together with generic identifications and affinities, substantiate this evidence and, in addition, show that the deposition of the Colpitts may have begun during the Ludwigian age and surely terminated before the end of the succeeding Sonninian age.

The ammonite species are largely new as are many of the genera, though some forms are identical with the restricted "genera" of Buckman, and others show close affinity to certain of his "species." *Tmetoceras* n. sp., cf. *T. scissum* (Benecke) occurs near the base of the Weberg formation. Associated with this species, in its upper range, are poorly preserved specimens of Strigoceratidae, apparently *Deltostrigites* Buckman, and also species of Sphaeroceratidae belonging to a new genus having the general external appearance of *Emileia*. Here is conflicting evidence regarding the age of the basal Colpitts. The close similarity between the *Tmetoceras* species and *T. scissum* (Benecke) suggests early Ludwigian age but the other species associated with it are indicative of the early Sonninian age. Stratigraphically above the *Tmetoceras* beds are species of Hebetoxyitidae, probably referable to *Hebetoxyites* Buckman, of which four species are defined and placed in the Sonninian age (Buckman, 1909-1930). Higher stratigraphically, in the upper part of the Weberg formation, comes the profusion of Sonninidae in which are represented some important new stages of the group. With few exceptions the Sonninidae of other parts of the world are retrogressive, showing particularly a loss of spines or costae during ontogeny. Many of the Colpitts species show marked progressive stages in which costae, nodes, and spines are acquired after a smooth stage. Others are in a secondary stage of progression showing spines and costae on the inner whorls and outer whorls separated by a smooth or slightly costate stage. Still others appear to be the normal retrogressive type of Sonninidae, but some of these may actually be in a secondary retrogressive stage in which earlier retrogression is obscured by poor preservation, by concealment of the inner whorls, or by elimination from the ontogeny. Many individual specimens are large, some as much as 14 inches in diameter. However important these new forms may be in adding unknown chapters to the

phylogeny of the Sonninidae, they are of no great immediate assistance in accurate dating of the beds in which they occur. Many of the species show marked resemblance in whorl shape, degree of coiling, and ornamentation to the early Sonninidae described by Buckman (1887-1907) under the name of *Sonninia* but later referred to the new genus *Euhoploceras* (Buckman, 1913) so as to set them off from the more retrogressive Sonninidae of later date. If the Oregon forms are as close in time to *Euhoploceras* as the above noted characters suggest, then they may be of "early" Sonninian age. However they have diverged so far from *Euhoploceras* in the ontogenetic arrangement of those characters that there is no assurance that they are not actually farther removed in time from *Euhoploceras* than *Sonninia s. s.* and its allies which occur in the "*Otoites sauzei* zone" of late Sonninian age.

Other ammonites associated with the Colpitts Sonninidae have more definite age significance. They include:

Sphaeroceratidae
Docidoceras, several new species.
Strigoceratidae
Deltostrigites n. sp., cf. *D. deltotus* Buckman.
Family?
Zugophorites, several new species.
Zugella, several new species.

The Warm Springs formation contains numerous ammonites, but most of the species belong to a new genus of uncertain affinities. Occasional limestone lenses contain large sonninids similar to those of the upper Weberg formation suggesting that the Warm Springs formation is not much later than the Weberg. The differences between the faunas of the two formations are probably in large part the result of facies differences.

The absence of Stepheoceratidae from the Colpitts fauna is of considerable significance. Stepheoceratid species are to be expected in faunas of late Sonninian age, and they become the dominant ammonite family in the next or "Stepheoceratan age." They were not absent from North America, for they have been reported from several localities in British Columbia and California where fossils are not abundant. Their absence from the rich faunas of the Colpitts group indicates that deposition ceased before the end of the Sonninian age.

Faunal evidence regarding the age of the Colpitts group may be summarized as follows: (1) The occurrence of *Tmetoceras* in the lowermost beds suggests that deposition might have begun during the Ludwigian age, but this evidence is made uncertain by the association of *Tmetoceras* with other species of early Sonninian affinities; (2) the affinity of all other identified forms mentioned above are with species whose chronologic positions in England are within what might be termed "early and

middle" Sonninian age; (3) none of the Oregon species has yet been identified with species which come from the well-known "*sauzei*" zone of "late" Sonninian age.

No faunas of early Sonninian age have been certainly identified elsewhere in North America, and this may explain the abundance of new species and genera in the Colpitts fauna. Faunas of slightly later date have been reported from British Columbia, Alberta, and California but they are probably in large part contemporaneous with the faunas of the Izee group.

IZEE GROUP

GENERAL FEATURES

The Izee group consists of two thick noncalcareous formations of the early Middle Jurassic which succeed the Colpitts group. The Hyde formation includes more than 1000 feet of massive blue-gray sandstones in the lower part of the Izee, and the Snowshoe formation includes about 2800 feet of shale and laminated sandstone in the upper Izee. This group is widely exposed along the Ochoco anticline from the western end of the Jurassic rock area northeastward across the South Fork Valley and into the west side of Bear Valley. It is also present in the undifferentiated Jurassic areas along the Silvies River and in the upper South Fork Valley.

The separation of the early Middle Jurassic deposits into Colpitts and Izee groups is made for several reasons. The Colpitts is present only in the western end of the Jurassic area whereas the Izee extends northeastward into the Silvies River region, beveling both Colpitts and Mowich groups and lying discordantly upon Triassic strata. There are also marked lithologic differences which are suggestive of geologic and climatic changes for, with the deposition of the Colpitts group, the sequence of thin limy formations ends, and the Hyde, Snowshoe, and even later formations are much thicker and largely noncalcareous. This change is emphasized by the marked transition from richly fossiliferous Colpitts beds to the generally unfossiliferous massive Hyde sandstone above. The faunas show only a small time interval between the two groups, but striking faunal differences indicate changed seaways and new routes of migration. Thus it would appear that the Colpitts and Izee represent two distinct sea invasions during the early Middle Jurassic.

HYDE FORMATION

Type area.—The type area of the Hyde formation is at South Fork Bridge, section 30, T. 17 S., R. 28 E., in the bottom of South Fork Valley. The Hyde of this area is almost entirely massive blue-gray medium-grained sandstone in which there is little differentiation into distinct

minor beds. A measured section on the high hill in section 30, T. 17 S., R. 28 E., immediately southwest of South Fork Bridge, showed a thickness of 1080 feet. This sandstone is a prominent lithologic feature that extends across the South Fork Valley and accentuates the topographic expression of the southeast limb of the Mowich anticline. Appearing from beneath the plateau lavas on the north side of Snow Mountain, it strikes northeast along the anticline and crosses the South Fork Valley almost at a right angle to the general course of the river. Most of the way, especially on the southwest side of the valley, the sandstone forms a distinct strike ridge which is interrupted in places by canyons of superposed streams tributary to the South Fork River. The sandstone has caused a marked constriction in the South Fork Valley, and the river cuts through it at South Fork Bridge in a narrow gorge. Northeast of the Bridge, the sandstone crosses lower Poison Creek and continues to the northeast along upper Rosebud Creek toward the Silvies River divide.

Some erosion of the older Jurassic formations before the deposition of the Hyde in the South Fork area is indicated by the absence of Colpitts group and, in places, of the Mowich group, so that the Hyde lies upon truncated Triassic beds. However, the presence of the thin Mowich group between the Hyde and the Triassic from Big Flat northeast to Poison Creek, a distance of 5 miles, shows that there was little folding of older Jurassic formations before the Hyde deposition.

The continuation of the Hyde formation northeast of the type area has not been fully studied, but it is apparently represented by the prominent sandstone unit which trends northeast from upper Rosebud Creek to Bear Valley Ranger Station along the west side of the uppermost Silvies River. The sandstone is especially well exposed along the Izee-Canyon City road for 2 miles south of Bear Valley Ranger Station.

Suplee-Snow Mountain area.—The Hyde exposures, between Snow Mountain and Suplee, lie on both sides of the Mowich anticline. They occupy irregular areas between the Colpitts exposures and the rims and tablelands of Tertiary lavas. The area on the south limb is the westward continuation of the type section of the South Fork area. The two areas are separated by the lava cap of Snow Mountain and Mowich Mountain which projects northward into the Triassic rock area between the limbs of the Mowich anticline. The south limb exposures occupy Howard Valley and the south and west margins of the Freeman Creek basin. They extend westward along the Ochoco Forest boundary, between the lava rim and the Colpitts group, to the headwaters of Warm Springs Creek. This section of the Hyde differs from that of the type section in the presence of about 200 feet of shale and fine sandstone at the base

and in the division of the blue-green sandstone into a number of distinct beds of massive sandstone separated by beds of softer fine-grained sandstone. The lower shaly part includes fine sandstone and shale which show various shades of green, yellow, and gray. The colors distinguish these beds from the underlying Warm Springs formation of the Colpitts group. The sandstone section above the shales includes some grit and pebbly sandstone at the base, but higher up it is largely bluish-gray and greenish-gray massive sandstone. Apparently no more than 1000 feet of the Hyde is exposed here. It does not conform to the steeply dipping Mowich and Colpitts groups on the south limb of the anticline in the Freeman Creek region. For more than 3 miles southward into Howard Valley, it occurs in gentle folds that contrast with the prevailing steep southerly dips of the older formations.

The Hyde formation is more widely exposed on the north side of the Mowich anticline than on the south side. Since it lies upon the Colpitts group except in its easternmost extension, the distribution of that formation previously recorded also roughly defines the distribution of the Hyde, though the latter occupies larger areas because of its greater thickness.

The principal north-side area lies in the syncline of Colpitts group on upper Camp Creek and extends northeastward to upper Pine Creek, though tablelands of Tertiary lavas and sediments cover the formation on both sides of the Beaver Creek canyon. A paced section on the east side of Camp Creek valley in the NE ¼ of section 31, T. 18 S., R. 26 E., gave the following sequence:

	Feet
Mostly fine- and medium-grained sandstone with occasional beds of sandy shale showing beneath the slope wash on the side of Windy Ridge below the Tertiary lava and sediments	770
Mostly concealed but showing sandy shale in places	115
Medium- and fine-grained massive bluish-green sandstone	130
Medium- and coarse-grained sandstone	180
Greenish grit and coarse sandstone	12
Greenish-yellow fine sandstone	20
Dark greenish-gray fine conglomerate: pebbles of limestone and chert mostly less than 1 inch in diameter; matrix of poorly sorted massive grit and sandstone	35
Green, yellow, and black shale with some fine-grained yellowish sandstone lying upon the Warm Springs formation of the Colpitts group	250
Total thickness exposed	1512

The structural relations of the Hyde and underlying Colpitts group present an interesting anomaly. The structures of the two formations are notably different in the syncline on upper Camp Creek and also in the Freeman Creek area on the south limb of the Mowich anticline. As

these two areas were among those studied during the first seasons of reconnaissance work on the central Oregon Jurassic, the anomalous relations were taken as indication of angular discordance. Later work revealed no place where the Hyde extended across truncated Colpitts group in the manner that would be expected if an angular unconformity were present. Though the Colpitts is absent east of Snow Mountain and the Hyde lies upon both the Mowich group and Triassic beds, no appreciable angular discordance with the Colpitts is indicated.

At least four factors have played a part in producing the different Colpitts and Hyde structures on Camp Creek and Freeman Creek: (1) The resistant Hyde sandstones are separated from the resistant Weberg formation of the lower Colpitts by over 400 feet of shale and fine sandstone in the upper Colpitts and lower Hyde, and much of this zone is obscured under soil-covered depressions; (2) much of the adjustment to folding has taken place in these shaly beds so that the Hyde sandstones above are less deformed than the lower Colpitts; (3) large intrusions of basalt porphyry have entered this zone in places and deflected the resistant beds from their former relative position; and (4) some of the small faults common in the Jurassic terrane are largely concealed in this shale zone and other soil covered areas. These factors have combined on upper Camp Creek to produce a shallow eastward-plunging syncline of Hyde sandstones whose outcrops are widely divergent from those of the steeply inclined Colpitts strata below. On the ridge west of Freeman Creek, the Mowich and Colpitts groups on the south limb of the Mowich anticline strike eastward and dip southward at more than 45 degrees. The Hyde formation above dips at a lower angle and lies in a southwest-plunging syncline so that the outcrops swing southeast and south away from the Mowich anticline. No Warm Springs shale and sandstone (upper Colpitts) could be found beneath the Hyde on the ridge top, but these are fully exposed on both sides of the ridge. Dips recorded for the Colpitts in this area range between 40 and 60 degrees, and for the Hyde they range from 10 to 40 degrees. The origin of these structural relations so suggestive of angular discordance cannot be adequately explained until a detailed map of the area is made, but their value as evidence for angular discordance between the Colpitts and Hyde is believed doubtful.

The reasons for referring the green, yellow, and gray shales and sandstones of the Suplee area to the basal Hyde instead of the underlying Warm Springs formation, thus placing the contact of the formations near the middle of a 400-foot section of fine sandstones and shales, are (1) the presence above this contact of green sandstone of the Hyde type interbedded with the shales, (2) the presence of stepheoceratid ammonites at

the base of the part referred to the Hyde formation, and (3) the above-mentioned possibility of angular discordance between these two units.

Age of Hyde formation.—Shell fragments are common in the massive sandstones of the Hyde formation, but identifiable fossils are rare. The only fossils that are of value for age determination come from the Suplee-Snow Mountain area. One collection came from sandy shales at the base of the formation 100 yards southeast of the Davis ranch house on the west side of Freeman Creek basin. In it are stepheoceratid ammonites, of the kind that Buckman refers to *Skirroceras,* and advanced sonninid ammonites similar to those of the "*sauzei* zone" and later deposits of Europe. Another collection comes from the basal grit of the Hyde sandstones in the NE ¼ of section 13, T. 18 S., R. 25 E., on the north limb of the Mowich anticline. It also contains stepheoceratid and advanced sonninid ammonites. The Hyde fauna differs markedly from that of the underlying Colpitts group, but is so closely allied to that of the overlying Snowshoe formation that the chronologic significance of two faunas can be treated together. For reasons set forth later under "Age of Snowshoe formation," these stepheoceratid and sonninid ammonites of the Hyde formation are believed to indicate the "late" Sonninian age of the early Middle Jurassic.

SNOWSHOE FORMATION

Type area in South Fork Valley.—The remaining thickness of the Izee group, approximately 2800 feet, is here termed the Snowshoe formation. Exposures of this formation are poor, but occasional outcrops show all gradations from laminated dark-gray sandstones to black shales. Flinty black concretions are fairly common and some have yielded fossils. The type area is located on the southeast limb of the Mowich anticline where the exposures lie in a narrow belt extending from Snow Mountain northeastward across South Fork Valley. The formation lies southeast of and parallel to the ridge of Hyde sandstone but is eroded to lower levels and largely covered by soil and vegetation. The Suplee-Izee road follows the base of the Snowshoe, near its contact with the underlying Hyde, as it descends the southwest side of the valley. The formation crosses the river between Poison Creek and Rosebud Creek, passing beneath the Izee school and grange hall.

One exceptional locality in the South Fork area is the only place where both Snowshoe and Trowbridge fossils have been found in sufficiently close stratigraphic proximity to determine, within reasonably accurate limits, the contact of these two similar formations. This locality is on Flat Creek, a creek that flows eastward along the north base of Snow

Mountain to join the South Fork River near Izee. On the east side of Big Flat it leaves the Triassic area within the eroded Mowich anticline, and begins its steep descent to the South Fork along a canyon that exposes successively higher beds of the Hyde and later formations. The Hyde sandstone and some associated porphyry intrusions confine the creek to a narrow gorge at the east edge of Big Flat, and southeast of the gorge are the fine sandstones and shales of the Snowshoe formation and the black shales of the Trowbridge. The strike of the Snowshoe beds, though variable, is generally north and northwest, showing pronounced deviation from the northeast strike prevailing elsewhere. Several characteristic Snowshoe ammonites were found in slabby sandstone near the middle of the formation. Farther downstream, where the first spurs come down to the creek from the lava rim of Snow Mountain, both Snowshoe and Trowbridge fossils were found. The contact of the two formations is near a fine-grained green sandstone bed about 100 feet thick. About 50 feet stratigraphically below this bed, a characteristic Snowshoe ammonite was obtained from a concretion in sandy shale. Above the green sandstone begins the thick Trowbridge shale section. Large concretions in this shale contain characteristic Upper Jurassic macrocephalitid ammonites, and one of these ammonites was found only 300 feet stratigraphically above the green sandstone. This indicates that the green sandstone may be the basal bed of the Trowbridge shale.

Suplee-Snow Mountain area.—The Snowshoe formation extends westward from the type area along the south limb of the Mowich anticline, but very little of it is exposed west of the Snow Mountain lava cap because the low dips, as indicated by the underlying Hyde exposures, and the close proximity of plateau lavas to the Mowich anticline, effectively conceals all but the lower part of the Izee group.

North of the Mowich anticline, however, the Snowshoe shales and fine sandstones lie in a syncline adjoining the Mowich anticline. The exposures occupy an area of several square miles and extend from the Beaver Creek Canyon northeastward to upper Pine Creek, where the syncline dies out.

Seneca area.—The Snowshoe formation is present 4 miles south of Seneca where the Silvies River leaves the upper canyon of the Silvies River and enters the north side of the Silvies Valley downwarp. Because of the difficulty of delimiting the Snowshoe from the overlying Trowbridge shale, the only beds that can be referred to the Snowshoe are those exposed in the vicinity of two very prolific fossil localities. One locality is 0.7 mile north of the Jack Craddock ranch house, in section 24, T. 17 S., R. 31 E., where a quarry for highway ballast exposed fresh rock

in 1929. The other locality is a mile northwest of the Craddock house, in section 26, T. 17 S., R. 31 E., where shales are well exposed in a railway cut. Fossils are undoubtedly abundant in other parts of the section, but there are few good exposures in the area. The lithology of the Snowshoe beds here does not differ materially from that of the type area except that the shales show some alteration by large intrusions of basalt porphyry.

Age of Snowshoe formation.—The faunas of the Snowshoe formation are made up mostly of two characteristic families of early middle Jurassic ammonites, the Sonninidae and Stepheoceratidae. Species of the two families are found together in several localities. The time ranges of the Sonninidae and Stepheoceratidae overlap in Europe during the late Sonninian and early Stepheoceratan ages, so their association in the Snowshoe formation indicates a correlation of that formation with the middle part of the Bajocian stage of Europe. Consideration of genera and species bears out this indication.

The sonninids are very different from the sonninids of the Colpitts group, for they are advanced types mostly identifiable with European genera. The stepheoceratids are early members of the family and closely related to European species. A collection from the lower Snowshoe at South Fork Bridge comes from concretions in shale about 200 feet stratigraphically above the base of the formation. In it is an evolute, ornate stepheoceratid ammonite which closely resembles *Skirroceras leptogyrale* Buckman of the "late" Sonninian age. Two sonninid species identified here belong to *Sonninites* Buckman; one is similar to *S. simulans* Buckman, and the other resembles *S. felix* Buckman. The species found in the middle and upper parts of the formation in the canyon of Flat Creek also include *Sonninites*. This genus appears only in the late Sonninian age in Buckman's chronology, though only four species are recorded there. Two species of *Sonninites* were reported by McLearn (1926) in the middle sedimentary division of the Hazelton group of British Columbia, for which a late Sonninian age was suggested. Strata yielding either stepheoceratid or sonninid ammonites at several other localities in British Columbia, Alberta, and California have been referred to the later part of the Sonninian age and early part of the Stepheoceratan age by McLearn and Crickmay.

Fossil localities of the Snowshoe formation on the north side of Silvies Valley have yielded abundant ammonites in addition to a few pelecypods. In a railroad cut a mile west of the Craddock ranch house the following ammonites were found:

Stepheoceratidae
Skirroceras cf. *S. leptogyrale* Buckman.
Sonninidae
Several species of *Papilliceras* Buckman including *P. stantoni* Crickmay and *P.* cf. *P. blackwelderi* Crickmay.
Hebetoxyitidae
Several species showing affinities with *Hebetoxyites hebes* Buckman and *H. clypeus* Buckman.

One mile east of this locality and 0.7 mile north of the Craddock ranch house, numerous specimens of stepheoceratid ammonites and a few sonninids were found. The stepheoceratids are mostly referable to the genus *Skirroceras* Mascke. One species is similar to *S. leptogyrale* Buckman.

All the genera listed above appear in the late Sonninian age. The known species are not sufficiently numerous to establish the true time range of each genus, but the presence of three genera whose known species are mostly confined to the late Sonninian age strengthens the probability that the Snowshoe fauna is of that date. Moreover, the date of *Skirroceras leptogyrale*, which bears such close resemblance to Snowshoe species, is given by Buckman as "Sonninian, *sauzei*." *Papilliceras stantoni* and *P. blackwelderi* Crickmay are described from the Mormon formation of Taylorsville, California, and these species are assigned by Crickmay (1933, p. 899) to hemerae in the late Sonninian age.

Thus the ammonites identified in these several collections coming from the lower, middle, and upper parts of the Snowshoe formation in three widely separated areas all show affinities with European forms reported in the late Sonninian age, and with American forms for which a similar date has been postulated. Furthermore, it must be remembered that the Snowshoe overlies the Hyde formation, that shales referred to the lower Hyde contain stepheoceratid and sonninid ammonites similar to the Snowshoe species, and that the Hyde is underlain by the Colpitts group which contains a rich fauna, of "early and middle" Sonninian aspect, in which no stepheoceratid ammonites have been found.

The evidence favors the conclusions that deposition of the Hyde-Snowshoe sequence did not begin earlier than the late Sonninian age and that it was deposited by the same sea that carried faunas of late Sonninian and Stepheoceratan ages into British Columbia, Alberta, and California. There is no direct evidence in the species collected from the Hyde and Snowshoe formations to indicate that deposition continued after the Sonninian age, though it is probable that even the restricted "genera" identified have a much greater time range than that indicated by the few known species. However, ammonites representing the genera *Epalixites* Mascke and *Stemmatoceras* Mascke have been found in a conglomerate at the base of the sandstone sequence that crosses the upper canyon of the Silvies River immediately south of Seneca. These specimens apparently have been reworked from the Snowshoe formation.

Both genera occur in the Stepheoceratan age of Europe, and McLearn (1928) has reported *Stemmatoceras* in the lower Fernie of Alberta. Therefore, it is reasonable to assume that the Snowshoe deposition continued into the Stepheoceratan age, and so the Hyde-Snowshoe deposition must be referred to some place within the time included by the "late" Sonninian and "early" Stepheoceratan ages of Buckman's chronology. Strata of this date in Europe are referred to the *Otoites sauzei* and *Teloceras blagdeni* zones in the upper Bajocian stage of the Middle Jurassic.

TROWBRIDGE SHALE

TYPE AREA

The type area of this formation trends northeast across South Fork Valley along the Mowich anticline where the beds dip steeply toward the southeast in apparent concordance with the underlying Snowshoe formation. The thickness of the formation is about 4000 feet, and consequently the exposures are limited to a belt that is about a mile wide. The shale emerges from beneath the plateau lava on the north side of the Snow Mountain fault block and trends northward across Flat Creek canyon. North of the canyon the shale passes beneath outliers of plateau lavas on Hack Hollow Buttes and reappears on the side of the main valley of the South Fork. In the bottom of the valley, the base of the formation crosses the South Fork road 0.4 mile below Rosebud Creek, and the top of the formation is at the base of the first conglomerate bed of the Lonesome formation 0.4 mile above Antelope Creek. Northeast of the river the Trowbridge shale lies mostly in the valley of Antelope Creek and the valley of upper Lewis Creek. The Izee-Bear Valley road follows this belt of shale outcrops most of the way to the Izee Summit.

The exposures show dark-gray and black shale with only minor amounts of arenaceous material. Large black flinty calcareous concretions, characteristic of this shale, are found even where the shale is concealed beneath a soil mantle. They occur mostly in bands ranging from occasional subspherical concretions along a bedding plane to a series of closely spaced lenticular concretions and concretionary beds, which grade into fairly even beds of the same flinty material. A few concretions have yielded all the known Trowbridge fossils of the South Fork region. Good exposures of the shale member on Flat Creek show an increasing amount of fine-grained greenish-gray sandstone beds mostly less than a foot thick in successively higher beds of the upper part of the formation.

The top of the Trowbridge shale contrasts sharply with the thick basal conglomerates and sandstones of the overlying Lonesome formation,

but the base of the Trowbridge is difficult to determine without the aid of fossils. The lithologic differences observed in the Flat Creek canyon, where the Snowshoe-Trowbridge contact was restricted to a stratigraphic section of 450 feet by middle and upper Jurassic fossils, are present in other parts of the South Fork region. The differences are (1) the presence of occasional beds of soft laminated sandstones in the Snowshoe as contrasted with the black pencil shales of the Trowbridge, and (2) the greater abundance and size of concretions and concretionary beds in the Trowbridge shales.

No attempt is made here to delimit Snowshoe and Trowbridge formations outside the type area, though their presence is indicated in various places by thick sections of shale and fine sandstone. As these formations are both soft and easily eroded, they are marked topographically by an elongate area of smooth rounded hills which lies at a lower level than the prominent ridge of underlying Hyde sandstone on one side and the higher hills of the overlying resistant Lonesome formation on the other side. Consequently the Snowshoe and Trowbridge formations form a conspicuous topographic feature where they cross the South Fork Valley, and this feature, together with occasional shale outcrops, continues on northeast of the Valley paralleling the headwaters of the Silvies River as far as the edge of the Bear Valley alluvium. The open "parks" in the timber along the upper Silvies River, such as Snowshoe Camp and Wickiup sage flats, are upon this shale section. Exposures of these formations also form the smooth low hills on lower Lonesome Creek and lower Corral Creek in the upper South Fork Valley, being repeated on the southeast limb of a syncline which adjoins the southeast side of the Mowich anticline.

AGE

An early Upper Jurassic age for the lower and middle parts of the Trowbridge shale is demonstrated by macrocephalitid ammonites found in Flat Creek Canyon immediately north of Snow Mountain. Some of these ammonites are referable to the genus *Lilloettia* Crickmay (1930) which has been reported only from the Harrison Lake region of British Columbia. Crickmay described only two species of this genus and states that they are "of slightly later date than most of the group (Macrocephalitidae) and corresponding in age to *Catacephalites*." It is not clear whether he refers to the genus *Catacephalites* or to the hemera of the same name in Buckman's chronology, but a date within the Proplanulitan age is implied. Fragmentary specimens associated with *Lilloettia* in the Trowbridge shale have the general aspect of gowericeratid ammonites.

The Macrocephalitidae and Gowericeratidae are confined to the lowermost Upper Jurassic in Europe. Representatives of these families are reported at several places in Canada in addition to the Harrison Lake locality. Several macrocephalitid species from the upper Fernie formation of the Blairmore region, Alberta, were described by Buckman (1929) who discussed also the chronologic significance of the family in detail and showed the known range as the entire Macrocephalitan age and most of the Proplanulitan age. However, he reached no definite conclusion regarding the date of the Blairmore species. Gowericeratids have been reported also from the upper part of the Yakoun formation (McLearn, 1929), from the upper Fernie formation (McLearn, 1928), and from the Minabariet formation (Crickmay, 1930). Crickmay (1930) states that other faunas, containing either macrocephalitids or gowericeratids, which have not been described or have been misidentified, are present on the Alaska Peninsula and in the Sweetgrass Hills, Montana. Livingston (1932) has found macrocephalitids in the upper Snake River Canyon near Mineral, Idaho. These faunas show a widespread sea over parts of western North America during the early Upper Jurassic, and it is distinctly separated in time from the earlier and later sea invasions of the Bajocian and the Argovian. The Trowbridge shale, therefore, represents an additional record of this early Upper Jurassic sea. Crickmay (1931, p. 68), assigns the faunas of this sea to the Proplanulitan age. McLearn, who has been the principal investigator of the faunas, notes their affinities with late Macrocephalitan and Proplanulitan forms of Europe, perhaps implying that they are of the same date, but stating specifically only that they are of early Upper Jurassic age. The writer is inclined to agree with McLearn.

LONESOME FORMATION

TYPE AREA

The type area of this formation extends from the Spoon Creek region on Snow Mountain northeast across the South Fork Valley along the axis of a syncline which adjoins the southeast side of the Mowich anticline. The Lonesome formation contains many resistant beds of conglomerate and sandstone, and consequently its exposures are marked by an elongate area of high hills and ridges which are bounded on either side by the low rounded shale hills of the underlying Trowbridge and Snowshoe formations. In the bottom of South Fork Valley, the base of the formation is 0.4 mile above the mouth of Antelope Creek on the northwest limb of the syncline, and on the opposite limb it is 0.1 mile above the mouth of Lewis Creek. The uppermost beds exposed are along the axis near Izee Post Office.

The base of the formation is marked by thick beds of sandstone and conglomerate which contrast sharply with the underlying Trowbridge shale. The amount of black shale in the formation as a whole exceeds that of the sandstones and conglomerates, but in its lower 500 feet the coarse beds equal or exceed the amount of shale. The lower part of the section, in the bottom of the South Fork Valley, shows three conglomeratic beds, each about 100 feet thick, alternating with black shale beds of equal thickness. The conglomeratic beds range from coarse sandstone to fine conglomerate. The coarse material is largely varicolored chert in poorly rounded to subangular pieces. The amount of shale increases above these lower beds so that it constitutes about three-fourths of the formation. Interbedded with the shales are many sharply defined beds of massive sandstone ranging in thickness from a few inches to more than 75 feet. The sandstones are gray, bluish gray, and greenish gray, medium- and fine-grained. Like the Hyde sandstones, they are very hard, and the thicker beds form distinct rocky ridges and spurs on the valley sides. The Lonesome beds dip at angles of 80 to 90 degrees in the South Fork syncline and some are overturned. The thickness of the formation, in the section along the South Fork River, is about 4000 feet.

AGE

No fossils have been found in the Lonesome formation, and stratigraphic evidence shows conclusively only that it is later than the Trowbridge shale of the early Upper Jurassic and earlier than the Upper Cretaceous beds of the Bernard Ranch locality (Packard, 1928). The coarse basal beds of the type Lonesome suggest an unconformable relationship with the earlier formations of the Jurassic. The cherts in the basal grit and conglomerate are similar to the Paleozoic cherts of the Suplee region which are so abundantly incorporated as clastic material in the Triassic formations and to a lesser extent in the Jurassic formations as far up in the section as the Hyde sandstones. The recurrence of conglomerates with poorly rounded chert pebbles, after deposition of 6000 feet of fine sandstone and shale (Snowshoe and Trowbridge formations), suggests that some diastrophic change, preceding the Lonesome deposition, caused near-by erosion of the lower part of the Jurassic section or of the Triassic and Paleozoic formations.

UNDIFFERENTIATED JURASSIC OF THE SENECA REGION

Differentiation of formations in the eastern part of the Jurassic area has not been completed. Their exposures lie along the south side of Bear Valley and continue southward along both sides of Silvies River

as far as Gold Hill. These deposits consist largely of shale with a lesser amount of blue-gray and greenish-gray massive sandstones and conglomerates. No fossils or strata resembling those of the very distinctive Mowich and Colpitts groups have been found in this section, and it is probably made up entirely of the post-Colpitts noncalcareous formations.

The beds are involved in a series of overturned isoclinal folds in which the prevailing dip is northward at 60 to 90 degrees. The formations are not so distinctive in their lithology as they are in the South Fork Valley, largely because the Hyde formation is not recognizable in the Seneca region as a thick, massive sandstone unit, but instead appears to be represented by nearly equal amounts of shale and sandstone. Therefore, it is similar to the much later Lonesome formation and also difficult to delimit from the overlying Snowshoe formation. There is also the possibility that later strata than the Lonesome formation are involved in the folds. The Seneca section differs from the South Fork Valley section also in the absence of the pre-Izee calcareous formations which would mark the bottom of the noncalcareous section as they do on the Mowich anticline. Erosion of the isoclinal folds has not been sufficient to expose the pre-Izee formations and consequently the repetition of noncalcareous strata is not complete. The Seneca section has been injected by a number of large dikes and sills of basalt porphyry which are as much as 3000 feet wide. Though diligent search has been made, fossils have been found in only a few places, due to the scarcity of good exposures. The shales, which contain most of the fossils, are largely obscured by soil and vegetation, and the good exposures show mostly massive unfossiliferous sandstone and basalt porphyry. Recent construction of a highway and a railroad through the canyon between Seneca and the Silvies Valley made only two fresh exposures in shale, but both of these yielded abundant fossils.

The thick section of nonresistant shale and fine sandstone of the Snowshoe and Trowbridge formations is repeated several times in the Seneca region. One section is exposed between lower Flat Creek and lower Jump Creek on the west side of Silvies Valley. Another broad shale area, probably including a fold on which the shales are repeated, lies around the north side of Silvies Valley between Camp Creek and Soda Mountain. Abundant fossils found here prove the presence of the Snowshoe formation. A fourth shale section trends eastward across Soda Valley. Soda Valley is little more than a small basin in the upper Silvies Canyon, and it owes its origin to the easy erosion of the shale. Although these repeated shale sections should be reliable in indicating the middle parts of the Izee-Trowbridge sequence, there are too many uncertainties to permit further attempt at recognizing formations in the Seneca region.

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The State College of Washington, Pullman, Wash.

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EOCENE ANAUXITE CLAYS AND SANDS IN THE COAST RANGE OF CALIFORNIA

BY VICTOR T. ALLEN

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ABSTRACT

Eocene anauxite clays and sands containing andalusite occur at Tesla, Brentwood, and the Potrero Hills in the Coast Range as well as under the Great Valley at Rio Vista and McDonald Island. The fossil evidence suggests that these sediments were deposited in Meganos D, Capay, and Domengine stages of the Eocene. The type Capay in the Coast Range contains no anauxite or andalusite. Therefore, it is proposed herein that the anauxite sands and clays of the Coast Range were deposited by Eocene streams flowing across the area which is now the Great Valley and deriving sediment from the same source in the Sierra Nevada that furnished the anauxite sands of the Ione formation of Capay stage. The anauxite sediments of the Coast Range and Great Valley areas were mixed with sediments derived from a Coast Range source.

INTRODUCTION

In connection with the study of the Ione formation of the Sierra Nevada foothills in 1928 anauxite sands and clays similar in mineral composition

to the Ione sediments were found by the author in the Coast Range at Tesla in a formation then considered as belonging to the Domengine division of the Eocene. At that time the fossils collected at the type locality suggested that the Ione formation was Meganos Eocene, and the structures indicated that the Ione represented deltas of Eocene streams which deposited auriferous gravels of the same mineral composition on the slopes of the Sierra Nevada and sediments in the area now the Great Valley of California. Because the writer was called away from California in 1928 it seemed best to limit the published discussion (Allen, 1929) of the Ione formation to the occurrences on the east side of the Great Valley including the Clements Well. Investigation of the possible westward continuation of these deposits in the Coast Range was left for further study. In 1938 while the writer was visiting Professor of Geology in the Summer Session of the University of California, Berkeley, he revisited certain Coast Range localities and discussed with California paleontologists the new division of the Eocene in California, the Capay, between the Domengine and Meganos divisions. Recognition of the Eocene Capay stage (Merriam and Turner, 1937) and assignment to it of the type Ione and the formation at Tesla containing the anauxite sands and clays, indicated a possible correlation of the anauxite occurrences on both sides of the Great Valley. This provoked the questions of whether anauxite formed contemporaneously in two widely separated areas because of similar climatic conditions or whether the anauxite deposits of the Coast Range were the westward continuation of the stream deposits along the eastern margin of the Great Valley. This paper attempts to answer those questions and is based upon several weeks of field study and collecting during the summer of 1939 preparatory to several months of microscopic, chemical, and X-ray investigation of the sediments.

The term anauxite sand is used here for a sand with conspicuous plates of the pearly kaolin mineral anauxite (Allen, 1928) having an alumina-silica ratio of 1:3 which Ross and Kerr (1930, p. 164) have suggested forms an isomorphous series with kaolinite. In some sands the anauxite comprises as little as 2 or 3 per cent but in others it forms 50 or more per cent with quartz and feldspar constituting most of the remainder.

ACKNOWLEDGMENTS

Field collecting in 1939, four chemical analyses, and several X-ray patterns by Dr. John W. Gruner of the University of Minnesota were made possible by a Project grant from the Penrose Bequest of the Geological Society of America for which the writer is grateful. The writer is indebted to the Amerada Petroleum Corporation for samples of drill cores

at Rio Vista, to Mr. G. C. Jester of the Standard Oil Company for courtesies extended, and to Mr. T. W. Donlon of the Standard Oil Company for his assistance at Rio Vista. Mr. A. S. Huey gave useful information about the Tesla quadrangle, which he has recently mapped. Mr. Albert

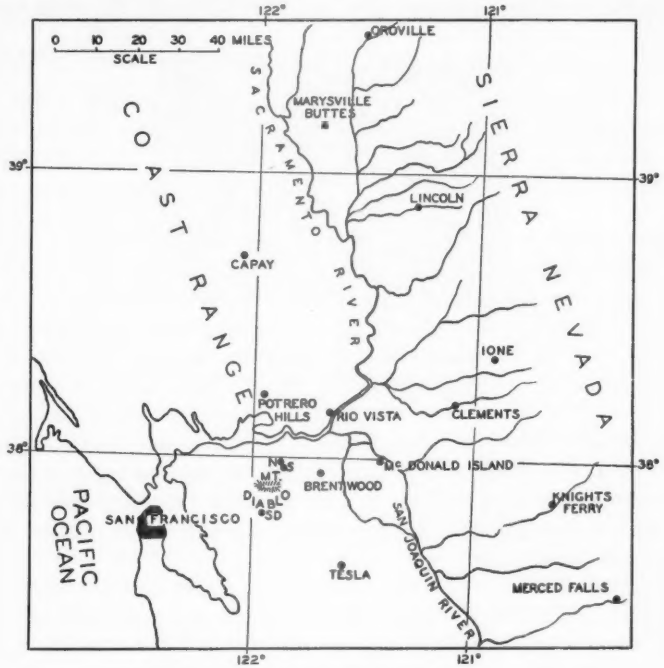

FIGURE 1.—*Index map of California*

Showing localities of anauxite and related sands. N—Nortonville; S—Sommerville; SD—South of Mt. Diablo.

Frank made mechanical analyses and prepared samples for petrographic examination. Dr. Bruce L. Clark identified the fossils collected during this study and has generously contributed time and helpful suggestions on the problem and certain areas familiar to him. The writer thanks Mr. L. Isobel of the Tesla Clay Company and other mine, pit, and well operators who granted permission to visit their properties. Finally, the author wishes to express appreciation to the members of the Department of Geological Sciences of the University of California, especially to Professor George D. Louderback and Professor Charles A. Anderson for help and encouragement.

TESLA AREA

Eocene clay and coal deposits at Tesla have been known and periodically mined since 1862. During a revival of mining activity in 1928 the writer visited the area and collected anauxite clays and sands (Loc. 1, Fig. 2) at the tunnel of the Livermore Clay and Sand Company 1¼ miles north-

FIGURE 2.—*Map of Tesla area*

Topography from U. S. Geological Survey, Tesla Quadrangle, California. Samples taken along line from 12 to 29.

west of Tesla. The mineral assemblages of these samples (No. 1, Table 1) were found to be remarkably similar to those of the Ione sediments on the east side of the Great Valley. In 1938 the author revisited and collected white anauxite clays (No. 3, Table 2) from Locality 3 and anauxite sands and clays from Locality 4.

In the summer of 1939 the tunnel of the Tesla Clay Company was visited (Loc. 2, Fig. 2) several times, and numerous samples of sands and clays were collected through the courtesy of Mr. L. Isobel. Among the samples were anauxite clays similar in composition to No. 2, Table 2 taken 170 feet from the shaft entrance. All the clay removed from this mine was reported to be suitable for the manufacture of pottery and was used successfully by a number of companies in the Bay district including the Technical Pottery Company, Oakland. In addition the section along the Tesla road was sampled (Nos. 12 to 29; Fig. 2).

EOCENE SECTION AT TESLA

Modified after A. S. Huey

		Feet
(29)	Light-gray biotite sand	8
(28)	White quartz sand, some layers containing anauxite (27)	40
(26)	Buff biotite sands	75
(25)	Gray clays interbedded with biotite sands	50
(24)	Fine buff micaceous sands	590
(23)	White clays and sands with gray shale	75
(22)	Thin-bedded anauxite sands and clays	8

	Feet
(21) Buff or iron-stained biotite sands, some concretions	300
(20) White anauxite quartz sand	20
(19) Dark brown shales; carbonaceous and lignitic shales	15
(18) Yellow iron-stained sandstone, some concretions	40
(17) Brown sandy shales with anauxite flakes and carbonaceous seams	150
(16) Micaceous sandstone and shale	300
(13) Thin layers of anauxite sand interbedded and in places mixed with biotite sands	40
(12) Micaceous gray and buff sands, some shale	240

Mr. A. S. Huey informed the writer that he collected *Turritella merriami* (Dickerson) near the top of the section in a buff sandstone similar to No. 24 in Table 1. This fossil also occurs at the Ione type locality and is the basis for assigning both the white clays from Tesla and from the Ione formation to the Capay.

The refractive index gamma of the white clays at Tesla is close to 1.566 ± .005, and gamma minus alpha is .006, so in optical properties as well as chemical composition (Table 2) the Tesla clays are similar to the anauxite clays near Ione. Dr. Gruner of the University of Minnesota, who made X-ray patterns of the white clays from Tesla and from Ione submitted to him by the writer, reports that all show identical anauxite patterns. The anauxite white sands and clays at Tesla both contain a heavy mineral assemblage similar to that in the Ione formation (Tables 1, 7). In some samples the minerals comprising the overlying and underlying beds occur with the minerals of the typical Ione assemblage. Some Eocene sands at Tesla contain much fresh black biotite, and a few contain sparse glaucophane, albite, and other typical Coast Range minerals. From the generalized Eocene section at Tesla it can be seen that biotite sands alternate at several horizons with anauxite sands and clays. Apparently these sediments were furnished from two different sources. As anauxite can form from biotite by loss of iron, magnesium, and other elements (Allen, 1928) fresh black biotite probably would not remain unaltered while other biotite in the same situation was altered to anauxite. Yet, biotite and anauxite occur in sample 18, Table 1. Furthermore, the leaching of biotite to form anauxite in a weathered granodiorite below the auriferous gravel at Nevada City (Allen, 1929, p. 387) was accompanied by the complete decomposition of the feldspars. Sample 13, Table 1, which is part of a 1-inch layer at Tesla overlain and underlain by biotite sands, contains over 50 per cent feldspars. The anauxite in sample 13 could not have formed from the decomposition of biotite after its deposition without change of the associated feldspars. Since deposition of sediments from a single source with pauses during the accumulation to allow leaching and alteration does not account for the sequence of sediments at Tesla, the derivation of material from two suggestive sources should be con-

TABLE 1.—*Minerals in Eocene sands at Tesla*

Number	Percentages										
	12	13	18	20	24	27	29	1	2	3	4
SIZE ANALYSIS (mm.): Greater 1				1.2							
1–1/2	7			7.5	2.4		4.8				
1/2–1/4	36			19.3	17.0		18.8				
1/4–1/8	34			35.8	44.6		41.6				
1/8–1/16	14			19.2	20.7		18.4				
less 1/16	9			17.0	15.3		16.4				
Total	100			100.0	100.0		100.0				
Minerals in 1/2 mm.–1/8 mm. separate: Quartz	73	41	76	60	73	72	74	86	79	57	89
Orthoclase	20	54	20	26	19	15	17	4	1	18	3
Plagioclase	4	0.7	2	6	3	1	2		0.5	1	0.8
Microcline		0.3		1					0.5		0.2
Biotite	3		1		5		7				
Anauxite		4	1	7		12		10	19	24	7
Green hornblende	C	D	E		C		C		E		
Blue-green hornblende	E	E									
Basaltic hornblende	E	E			E						
Glaucophane	E				E						
Actinolite	E										
Epidote	D	D	D	D	D	D	C	E	E	D	D
Sphene	D										
Zircon	D	E	D	D	C	D	D	C	D	D	D
Garnet	E	E	E	E	C	E	E	E		E	E
Rutile	E	E	E	E	E	E		E		E	E
Muscovite	E	E	E								
Andalusite	E	E	E	E	E	E	E	E	E	E	E
Tourmaline (brown)	E	E	E	D	E	E	E	D	E	E	D
Tourmaline (blue)		E	E	E	E	E	E	E	E	E	E
Zoisite				E	E						
Magnetite-ilmenite	D	D	D	D		D	D	C	D	D	D
Leucoxene	E	E	E	D	D	D	D	D	C	E	E
Percentage of minerals heavier than bromoform	2.0	1.5	1.2	0.8	2.0	0.9	0.3	0.3	0.4	0.2	0.3

B = abundant; C = very common; D = common; E = rare.

Position of samples 12, 13, 18, 20, 24, 27, 29 given in stratigraphic section and location of each shown in Figure 2.

Sample No. 1. White anauxite sand near tunnel of Livermore Clay and Sand Company, collected in 1928.

Sample No. 2. Anauxite sand Tesla Clay Co. tunnel (Loc. 2, Fig. 2).

Sample No. 3. Anauxite sand test pit (Loc. 3, Fig. 2).

Sample No. 4. Anauxite sand in gully (Loc. 4, Fig. 2).

sidered. In the Coast Range of California the Cretaceous and older rocks contain feldspars, biotite, amphiboles including glaucophane, and nearly all the other minerals, except andalusite and anauxite, which are present in these Eocene sediments. If erosion of rocks composed of these minerals

TABLE 2.—*Chemical composition of Tesla and Ione clays*

Number	2	3	4	5
SiO_2	48.39	47.90	48.93	48.00
Al_2O_3	35.77	35.91	36.13	34.56
Fe_2O_3	.91	.96	1.15	1.54
MgO	.23	.30	.21	.35
CaO	.10	.10	.36	.23
Na_2O	.09	.08	.87	.51
K_2O	.42	.42		.59
H_2O+	12.39	12.73	12.35	12.30
H_2O-	.95	.83		1.30
TiO_2	.53	.54		1.00
Ignition loss less total water	.27	.30		
Total	100.05	100.07	100.00	100.38

No. 2. White anauxite clay, Tesla Clay Co. tunnel 170 feet from entrance (Loc. 2, Fig. 2). R. B. Ellestad, analyst.
No. 3. White anauxite clay, test pit near Tesla (Loc. 3, Fig. 2). R. B. Ellestad, analyst.
No. 4. White clay, Ryan ranch, NW. of Tesla. C. W. Briggs, analyst (Dietrich, 1928, p. 354).
No. 5. White anauxite clay, Ione formation at Jones Butte, 3 miles NW. of Ione, California (Allen, 1929, p. 379.)

takes place so rapidly that weathering does not destroy them before deposition in a basin, the highly feldspathic, ferromagnesian members of the series could form. Because andalusite and anauxite do not occur in the underlying Cretaceous at Tesla, it is only natural to choose a source known to contain these minerals, namely the Sierra Nevada bedrock, which near Ione during the Capay supplied some quartz sands containing these minerals with less than 1 per cent of feldspars. All sediments at Tesla could be formed by material from either of these sources or by mixing the materials from both sources in the proper proportions.[1]

[1] Since this manuscript was prepared a similar occurrence of anauxite sands containing andalusite mixed with Coast Range minerals was found in a Domengine formation in the Panoche Hills, Fresno County, by Max Payne, who kindly sent samples to the writer for examination.

BRENTWOOD AREA

Professor Charles A. Anderson first called the author's attention to the andalusite in the glass sands at Brentwood. Sands (No. 36) with over 60 per cent anauxite occur above the lower glass sand at the Silica Company of California Mine, where the following composite section was assembled.

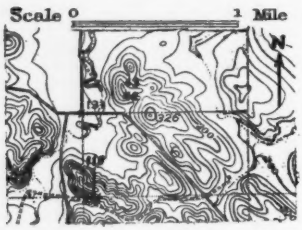

FIGURE 3.—*Map of Brentwood area*

Topography from U. S. Geological Survey, Byron Quadrangle, California. Samples taken along line from 161 to 115.

COMPOSITE SECTION OF MINE OF SILICA COMPANY OF CALIFORNIA

4 MILES SOUTH OF BRENTWOOD, CALIFORNIA

	Feet	Inches
(43) White glass sand with occasional plate of anauxite. Top of old glass sand pit		
(42) White glass sand with occasional pebble of quartz or plate of anauxite—middle of old pit of Silica Company of California. According to B. L. Clark these are in the Domengine division of the Eocene, being above the basal conglomerate		
Covered area		
Iron-stained clayey sands containing biotite	14	
Iron-stained sandy clay with anauxite and biotite	4	
Gray sands containing biotite and iron spots	6	
Hard buff calcareous sandstone		10
(115) Sand containing biotite, iron stains, and lenses of hard gray calcareous sandstone	5	
Iron-stained sandy clay poorly bedded, contains biotite	8	
(113) Thinly bedded anauxite sand	1	
Hard calcareous buff fossiliferous sandstone		14
Iron-stained clay and sand with lenses of anauxite sands	12	
Yellow sandy clay with interbedded chlorite and anauxite		20
(37) Gray calcareous fossiliferous sandstone. Contains *Turritella merriami* subsp. *brevitabulata*		18
Iron-stained sandy clay	21	
Hard calcareous sandstone	1	
Thinly bedded clayey sand, lenses of biotite sand (35) overlain by sands (36) with abundant plates of anauxite. (Analysis No. 1, Table 3)	6	
Iron-stained clayey sand	4	
Iron-stained quartz sand	5	
(159) Glass sand mined by Silica Company of California	10–15	
Iron-stained quartz sand	10	
(160) Anauxite sand	10–30	
(161) Iron-stained biotite sandstone		

Although equal care was given to hand picking and using the necessary heavy solutions on the anauxite from this locality and that near Ione, it was impossible to remove all the quartz and feldspars from the edges and between the plates of the Brentwood anauxite. The resulting

TABLE 3.—*Chemical composition of anauxite plates from Brentwood and Ione*

Number	1	2	3	4
SiO_2	60.45	55.0	52.46	55.88
Al_2O_3	20.25	31.1	32.20	30.24
Fe_2O_3	1.73	1.4	1.69	.58
MgO	.43	.41		
CaO	.80	.01	.03	tr.
Na_2O	.53	n. d.	0.25	0.34
K_2O	1.41	n. d.	0.31	0.42
H_2O+	7.34	12.13	12.07	11.72
H_2O-	3.19		1.38	.63
TiO_2	2.57	n. d.	0.55	0.50
MnO	.02			
Ignition loss less total water	1.30			
Total	100.02	100.05	100.94	100.31

No. 1. Anauxite plates with about 10 per cent quartz from Eocene sands above glass sands at Silica Company of California Pit (Loc. 36, Fig. 3) Brentwood. R. B. Ellestad, analyst.
No. 2. Anauxite plates, Ione formation, Mokelumne River. W. S. Morley, analyst (Allen, 1928, p. 147).
No. 3. Anauxite plates, Ione formation, Mokelumne River. J. G. Fairchild, analyst (Allen, 1928, p. 147).
No. 4. Anauxite plates, Ione formation, Mokelumne River. W. F. Hillebrand, analyst (Allen, 1928, p. 147).

chemical analysis given as No. 1, Table 3 is higher in silica and lower in alumina than the analyses of the Ione anauxite. Dr. Gruner, who made X-ray patterns of the separated sample, reports that it shows an X-ray pattern identical with that of anauxite plates from Ione, but in addition there is also about 10 per cent of quartz present. The results agree closely with those of the optical examination, suggesting that the high silica reported can be accounted for by grains of quartz and orthoclase and that the high potash is due in part to small grains of orthoclase and muscovite attached to the anauxite plates. Some anauxite plates from Brentwood are optically identical with those from the Ione formation

(Allen, 1928), but the refractive index gamma of some is lower than the 1.564 given as typical for Ione anauxite. Many plates show inclusions of rutile needles, and the 2.57 per cent titania reported by Ellestad is higher than that present in the Ione anauxite. It is somewhat disappointing that what to the writer's knowledge is the first reported analysis of anauxite plates from the California Coast Range should show contamination. However, contamination is perhaps to be expected if the anauxite originated in the Sierra Nevada and was transported by streams to the Brentwood area, where gypsum, muscovite, and about 10 per cent of orthoclase occur in a layer containing more than 60 per cent anauxite. Layers of Ione sand at the Newman pit, near Ione, with the same per cent of anauxite contain less than 1 per cent orthoclase. At Brentwood evidence for the derivation of these unusual Coast Range sediments from two sources is more clearly displayed than at Tesla. Below the anauxite sands are sands (No. 35, Table 4) with over 70 per cent biotite and with as much orthoclase as in the overlying anauxite sand. Some layers contain nodules having a center of anauxite plates completely encased by biotite that must have escaped alteration if weathering accompanied accumulation. A section perpendicular to the cleavage of one nodule (Pl. 1, fig. 5) shows the anauxite in white surrounded by a green platy mineral more nearly chlorite than biotite. Relatively, the green mineral has altered little when compared with the end product anauxite. Certainly any solutions that would form anauxite from biotite in the center would also attack the biotite along the top or bottom of the nodule. Therefore, it is difficult to explain this relationship by alteration of the biotite either during or after deposition, but all variations could have resulted from biotite washing in from one area and anauxite from another.

In all the biotite sands at Brentwood feldspars range from 10 to 40 per cent; the sample of glass sand (159) contained over 20 per cent. Interbedded with the biotite and anauxite sands at Brentwood are hard gray calcareous sandstones composed of abundant green amphiboles and biotite as well as nearly all the other minerals reported by Woodford (1927, p. 76) from the type Meganos. Among several fossils in a calcareous sandstone (No. 37) was one identified by Dr. Clark as *Turritella merriami* subsp. *brevitabulata* characteristic of the Meganos D division of the Eocene. Several thin sections were made of the upper oxidized part of the calcareous sandstones and compared with the fresh sandstone (Pl. 1, fig. 6) to see whether anauxite had formed from biotite during the weathering, but no new minerals accompanied the development of iron oxides. The contact of the hard calcareous sandstone with the overlying sand was sharp, and there was no indication that the

TABLE 4.—*Minerals in Eocene sands at Brentwood*

Number	Percentages								
	161	160	159	35	37	113	115	42	43
SIZE ANALYSIS (mm.): Greater 1				1.2				12.4	
1–1/2	3.7	23.6	39.0	27.8	.3	5.9	3.3	20.6	0.1
1/2–1/4	57.7	43.1	39.1	26.4	7.6	10.9	6.6	39.0	28.4
1/4–1/8	18.0	18.6	12.3	20.8	34.6	34.4	29.2	16.4	54.3
1/8–1/16	10.9	8.1	6.0	12.8	37.9	29.7	40.4	6.8	11.5
less 1/16	9.7	6.6	3.6	11.0	19.6	19.1	20.5	4.8	5.7
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0
calcite cement	28				40.8				
Minerals in 1/2 mm–1/8 mm. separate: Quartz	67.0	58.0	78	16	55	71	58	92	81
Orthoclase	30.0	31	21	11	42	22	34	7	18
Plagioclase	1	2	0.5	1	2	3	2		0.3
Microcline	1	2	0.5			1		0.5	0.5
Biotite	1			72	1		6		
Anauxite		7				3		0.5	0.2
Green hornblende	E	E		D	B	D	B		
Blue-green hornblende					E	E	E		
Basaltic hornblende					E	E			
Glaucophane						E			
Actinolite	E			E	E	E			
Epidote	C	E	C	C	D	C	C	C	D
Sphene	E			E	E				
Zircon	D	D	D	D	D	D	E	D	D
Garnet	E	E	E	E	E	E	E		
Rutile				E		E		E	E
Muscovite	E	E		E	E		E	E	E
Andalusite	E	E	E	E	E	E	E	E	E
Tourmaline (brown)	E	D	D	D	D	D		D	D
Tourmaline (blue)				E	E	E		E	E
Zoisite				E					
Magnetite-ilmenite		D	D	C		D	C	C	D
Leucoxene		D	D	E	D	D	D	E	E
Percentage of minerals heavier than bromoform	0.2	0.1	0.05	3	4	0.3	0.4	0.2	0.3

B = abundant; C = very common; D = common; E = rare.
Position of samples 161 to 43 given in composite section, and localities of each shown in Figure 3.

TABLE 5.—*Minerals in Eocene sandstones and sands*

Number	Percentages										
	48	51	45	170	153	152	64	52	55	163	71
SIZE ANALYSIS (mm.). Greater 1						6.7			5.1	7.0	
1–1/2			.3	6.8	6.0	32.0		2.6	34.9	11.2	9.1
1/2–1/4	5.6	54.5	35.9	5.4	8.3	32.3		62.3	34.5	32.7	48.6
1/4–1/8	55.0	32.1	42.9	8.4	35.3	14.5		20.8	13.9	21.2	17.1
1/8–1/16	21.6	7.2	12.2	56.4	32.3	7.8		8.1	8.2	14.5	15.6
less 1/16	17.8	6.2	8.7	23.0	18.1	6.7		6.2	3.4	13.4	9.6
Total	100.0	100.0	100.0	100.0	100.0	100.0		100.0	100.0	100.0	100.0
Minerals in 1/2 mm.–1/8 mm. separate: Quartz	77	87	74	72.5	75	86	79	78	82	50	95
Orthoclase	22	13	26	27	23	14	12	16	2	46	1
Plagioclase	0.5									3	
Biotite				0.5				2		1	
Anauxite	0.5				2		9	4	16		4
Green hornblende						E			E	B	
Blue-green hornblende										E	
Basaltic hornblende										E	
Glaucophane											
Actinolite										E	
Epidote		D		D	D	D	D		D	D	D
Sphene								E	E	E	
Zircon	C	C	D	D	C	C	C	D	D		D
Garnet			E	E	E		E	E		E	
Rutile	E			E		E					
Muscovite	E			E	E					D	
Andalusite					E	E	E	E	E		E
Tourmaline (brown)	D	D	D	D	D	D	D	D	D		D
Tourmaline (blue)	E				E	E			E		
Zoisite										E	
Magnetite-ilmenite	C	D	D	D	C	C	C	C	C	C	C
Leucoxene	D	D	E	E	D	D	D	D	D		D
Percentage of minerals heavier than bromoform	0.5	.03	0.7	0.15	0.2	0.2	0.2	0.2	0.2	2	0.9

B = abundant; C = very common; D = common; E = rare.

(48) Shaly sand, near loading bin, Domengine. Sommerville, Calif.

(51) Glass sand, Domengine. Sommerville, Calif.

(45) Glass sand, Domengine. Nortonville, Calif.

(170) White sand, Domengine, south of Mt. Diablo, Calif.

(153) Anauxite sandy shale, Potrero Hills, Calif.

(Notes continued on page 283)

anauxite sand formed by decomposition *in situ* of the calcareous feldspathic biotitic sandstone.

Sands (Nos. 42, 43) from an abandoned glass sand mine located across the road to the north (Fig. 3) contain anauxite and other minerals of the Ione assemblage. According to Dr. Clark these sands are above the basal Domengine conglomerate and belong to the Eocene Domengine division. The writer did not find the basal conglomerate at this locality, but it may have been present under the covered portion. Since the Capay is considered to follow the Meganos D division in other localities the question might be asked could the upper anauxite sands at Brentwood be Capay and the equivalent of the anauxite sands at Tesla and Ione? However, on the basis of his mapping Dr. Clark believes the upper sands are Domengine, the equivalent of the white sands at Nortonville and Sommerville.

All the anauxite and associated sands at Brentwood contain andalusite, which is lacking in the underlying Cretaceous and Martinez and is unknown in the older rocks of the Coast Range. The alternation of biotite sands with anauxite sands repeated in the section at Brentwood is suggestive of material being carried in from two sources, one in the Coast Range to supply feldspars, glaucophane, and blue-green amphiboles, the other a Sierran terrane to supply andalusite, blue tourmaline, anauxite, and other Sierran minerals. The presence of marine fossils in the calcareous sandstones shows that some of this material was deposited in the sea, but some of the upper anauxite sands at Brentwood have fluviatile bedding similar to that in parts of the Ione formation along the Sierran foothills thus indicating that at times the sea shallowed and stream characteristics predominated.

RIO VISTA AND McDONALD ISLAND WELLS

In order to ascertain whether anauxite sediments extend eastward from Brentwood and Tesla under the Great Valley to tie in with those of the Sierra Nevada, samples from the wells at Rio Vista and McDonald Island were examined. Anauxite sands with a mineral assemblage like that of the Ione sediments were found at a depth of 4710 feet in the Happe No. 1 well, Rio Vista (No. 55, Table 5) and at 4951 feet (No. 64, Table 5) in the McDonald Island well No. 1. Gray, white, or iron-stained clays and beds of lignite associated with the white anauxite

(152) Cliff sandstone, Potrero Hills, Calif.
(64) Anauxite sand, McDonald Island Well No. 1, depth 4951-4954 feet. G. H. Sturgeon collection.
(52) Biotite anauxite sand, Happe No. 1 well, depth 4571-4595 feet. Rio Vista.
(55) Anauxite sand, Happe No. 1 well, depth 4710-4737 feet. Rio Vista.
(163) Pebbly siltstone, type Capay, Capay Valley.
(71) Anauxite sand, Marysville Buttes.

sand in these wells resemble the Ione sediments at the type locality. Some of the clays have the optical properties of anauxite, but one from the D. D. well No. 1 at a depth of 4309 feet has refractive indices higher than those of anauxite but within the range of the indices of illite (Grim,

TABLE 6.—*Chemical composition of clay from well at Rio Vista and Ione clays*

Number	1	2	3	4
SiO_2	47.21	55.72	50.38	52.72
Al_2O_3	31.97	24.56	32.11	32.39
TiO_2	1.18		.38	
Fe_2O_3	3.65	4.02	2.99	2.45
MgO	.98	1.15	1.07	.39
CaO	.21	.45	1.15	.90
Na_2O	.39	1.18		.20
K_2O	1.30			.12
H_2O+	11.04	12.92	11.29	10.70
H_2O-	1.39			
Ignition loss less total water	.40			
Mn_2O_3	———	———	.56	
Total	99.72	100.00	99.93	99.87

No. 1. Gray clay from D. D. well at Rio Vista at depth of 4309 feet. R. B. Ellestad, analyst.
No. 2. Ione clay, Valley Springs. Analysis through courtesy of W. F. Dietrich (Allen, 1929, p. 379).
No. 3. Ione clay, Gladding McBean Pit, Lincoln. Analysis through courtesy of Company (Allen, 1929, p. 379).
No. 4. Ione clay. Lincoln Clay Products Co., Lincoln. Analysis through courtesy of W. F. Dietrich (Allen, 1929, p. 379).

1937). Dr. Gruner reports that the X-ray pattern of this clay is like that of anauxite with additional lines of another mineral. Its chemical composition (No. 1, Table 6) is suggestive of the Ione clays, somewhat high in iron, that occur at Lincoln and Valley Springs. The high potash content of this clay is also in keeping with the suggestion that it is a mixture of anauxite and illite. Some sands at Rio Vista contain both biotite and anauxite (No. 52, Table 5) as well as 15 per cent or more of feldspar, thus resembling the mixed types at Tesla and Brentwood.

Below beds containing fossils recognized as Cretaceous by William F. Barbat of the Standard Oil Company are white sands containing biotite extending to a depth exceeding 7000 feet below the surface at Rio Vista.

None of the Cretaceous white sands examined contained andalusite. The Cretaceous sediments at Pentz, near Oroville, and near Folsom also lack andalusite. The Eocene and younger sediments along the eastern margin of the Great Valley contain andalusite derived from the contact zones around the granitic intrusions in the Sierra Nevada. Does this indicate that contact aureoles, metamorphosed sufficiently for andalusite to have developed from carbonaceous sediments, were not exposed by erosion until Eocene time? If the lack of andalusite in the Cretaceous is consistent it might be a simple way of distinguishing Eocene from Cretaceous sediments in the Great Valley.

The well cores at Rio Vista and McDonald Island make it possible to trace anauxite sands containing andalusite under the Great Valley within 25 miles of those recorded at the Clements well (Allen, 1928).

POTRERO HILLS NEAR SUISUN

In the Potrero Hills which extend from 2 to 8 miles southeast of Suisun, Bailey (1930, p. 325) described a massive white sandstone as the Eocene Cliff sandstone. This sandstone (No. 152, Table 5) contains andalusite and other minerals of the Ione assemblage but no anauxite. However, in the underlying cocoa-colored shales are sandy shales with pearly plates of a clay mineral believed to be anauxite. This mineral crumbles more easily than most anauxite plates, and in general its refractive indices are lower than those of typical anauxite, but the range of indices is within those of the Brentwood material, which gives the X-ray pattern of anauxite. The Potrero Hills locality is the northernmost locality in the Coast Range where the Ione assemblage has been thus far discovered.

SOMMERVILLE, NORTONVILLE, AND SOUTH OF MT. DIABLO

At Sommerville and Nortonville, northeast of Mt. Diablo are white and gray sandstones that have been mined and used for the manufacture of glass and that Dr. Clark assigns to the Domengine (Clark and Woodford, 1927, p. 72). Although these sandstones resemble those with an Ione mineral assemblage the writer noted in 1926 that they lacked andalusite and anauxite. Several samples were collected in 1939 from different parts of the formation, and none (Nos. 48, 51, 45, Table 5) of those examined contained andalusite. One sandy shale contained a substance which is probably anauxite, but most of the clay mineral has refractive indices lower than those of typical anauxite.

While on an excursion conducted by Dr. Clark in connection with the Sixth Pacific Science Congress in 1939, the writer collected Eocene sandstones from a hill near the saddle between Pine and Curry canyons on

the south side of Mt. Diablo. Although these sandstones are unfossiliferous they rest on fossiliferous Cretaceous rocks and on the basis of lithologic similarity are considered equivalent to the Domengine or Capay white sandstones. The rocks examined (No. 170, Table 5) contain biotite but lack anauxite and andalusite.

Since andalusite is a resistant mineral, lack of it in these Eocene sands is difficult to explain if the sediments originated in the Sierra Nevada region which furnished the anauxite white sands. Clark and Woodford (1927, p. 78) have postulated a granite area to the east of Mt. Diablo now buried beneath the Great Valley. If such a granite area existed and lacked andalusite, the minerals of these sediments could have been derived from it, but these writers conclude that such a Great Valley granite is highly doubtful. Derivation of these Eocene sands from a Coast Range granite would require the removal of associated glaucophane and other characteristic Coast Range minerals. In forming this kind of sediment at the expense of easily weathered ferromagnesian and feldspathic materials, proof would be destroyed that Coast Range material had taken part in the process. Thus far the mineral composition of these Eocene sediments offers little direct evidence of their original source.

CAPAY VALLEY

On the basis of fossil evidence it appears that the deposition of anauxite sediments began in the Coast Range near the close of the Meganos and continued through Capay time into the Domengine. The type locality of the Capay was examined to see if the sediments contained anauxite deposited during that stage. The samples of pebbly siltstone containing the Eocene fossils collected in Capay Valley (Merriam and Turner, 1937) lack anauxite and andalusite (No. 163, Table 5) but are composed of abundant feldspars and ferromagnesian minerals (Pl. 1, fig. 3). Therefore, the conclusion seems justifiable that climatic conditions in the Coast Range were not favorable for the formation of anauxite and that no local supply of andalusite was available in this area.

MARYSVILLE BUTTES

On the south and west sides of Marysville Buttes narrow belts of anauxite sands are exposed. Williams (1929, p. 124-126) mapped these as a separate geologic unit on the basis of the writer's work (Allen, 1929) in the Sierra Nevada foothills and assigned them to the Eocene Meganos division because the anauxite sands lie between Meganos fossils in the underlying Marysville formation and the overlying Butte gravels. Because Merriam later found Cretaceous fossils in a calcareous sandstone

overlying the anauxite sands, some believe that there were anauxite sands in the Cretaceous at this locality. Others consider that the Cretaceous fossils have been moved from the underlying sediments and that the anauxite sands are Eocene. Suggested explanations include the following:

(1) The blocks containing the Cretaceous fossils were faulted into their present position among the Butte gravels, after the sedimentary series was tilted by the intrusion of the andesite. This faulting was connected with the later intrusion of rhyolite, so that slivers of Cretaceous rocks were forced upward along the strike of the Butte gravels. As shown by Williams' map (1929), faulting is evident on the west side of the Buttes, but in 1939 the writer could find little confirmation of strike faulting along the south side.

(2) The Cretaceous rocks were reworked during the Eocene so that Cretaceous fossils were washed out and redeposited in the Butte gravels. The writer examined Merriam's collection as well as the Cretaceous fossils he found among the Butte gravels but noted little suggestion of water-worn shells.

(3) Recently an oil company reported that in a trench across the Butte gravels Cretaceous fossils occur in blocks and boulders of calcareous sandstone which were deposited parallel to the strike of the Butte gravels.

Therefore, the latest evidence tends to return the anauxite sands at Marysville Buttes to the Eocene. The mineral composition of these sands (No. 71, Table 5) is similar to that of the Eocene Ione sands.

OTHER WHITE SANDS IN THE COAST RANGE

The writer does not consider all the white sands in the Coast Range to have come from the Sierra Nevada. The term white is relative and as used includes a great variety of sands, ranging in age from Cretaceous to Miocene. Some contrast slightly with the usually gray, buff, brown or iron-stained sands of the Coast Range but may differ from them only slightly in mineral composition. The origin of each type must await further knowledge; the one suggested in this paper applies only to the anauxite sands.

PROPOSED ORIGIN OF ANAUXITE SANDS OF THE COAST RANGE

This petrographic study indicates that anauxite is not a characteristic or common mineral in the Eocene sediments of the Coast Range since the type locality of the Capay in the Coast Range contains none. Nevertheless, deposition of anauxite did occur during Capay time at Tesla, Ione, and elsewhere. The sediments containing anauxite usually have some

grains of andalusite, but the underlying Martinez, Cretaceous, and older rocks of the Coast Range contain none. Any source furnishing anauxite should also provide andalusite.

In the Sierra Nevada during the Eocene the bedrock was covered with anauxite and other resistant minerals of a typical lateritic profile (Allen, 1929, pp. 383-394). Favorable topographic conditions and the warm moist Eocene climate caused the feldspars and ferromagnesian minerals to be changed to clay minerals. In the fresh and altered bedrock of this area andalusite occurs as a contact metamorphic mineral around the granitic intrusions.

During the Eocene the Sierra Nevada block was tilted westward, and streams began to erode the weathered products, carrying them westward across the area which is now the Great Valley. If we can rely on the fossil evidence, it seems probable that some of the materials reached the site of the present Coast Range in Meganos D time, forming the glass sands and anauxite sands and clays at Brentwood. Parts of the Coast Range area stood above the Eocene sea as islands and were eroded before a lateritic profile could form on them. Rapidly eroded surface materials were deposited in the same basin of accumulation, either as entire beds or mixed with minerals from the Sierran source.

Erosion continued during the Capay stage; and clays derived from the upper leached part of the lateritic profile were carried to Tesla, so that anauxite clays with very little flux and great refractoriness were deposited among beds coming from Coast Range erosion. Later in Capay time the shore line advanced eastward across the San Joaquin valley area to the foothills of the Sierra Nevada. The clays of the Ione formation, derived largely from parts of the profile lower than that which furnished the clays at Tesla, are not leached to the same extent but show the same kind of weathering. The deltas of the Ione formation now exposed in the Sierra Nevada foothills were formed mainly at this time.

Further uplift shifted the shore line westward. Partial erosion of the Ione sediments and the bedrock furnished anauxite sands in the Domengine.

A few California geologists think the Eocene white sands of the Coast Range are too angular to have come from the Sierra region. For comparison, representative grains between .5 mm. and 1 mm. from Tesla and Brentwood were photographed (Pl. 2, figs. 3, 4) and placed alongside photographs of similar-sized grains from the Calaveras River, collected 30 miles apart. It appears that these Coast Range sands are no more angular than the grains of modern stream sands. Even grains as large as 1 mm. which have travelled an additional 30 miles downstream are not noticeably different in rounding from those taken nearer the source

upstream. (*Compare* Figures 5 and 6 of Plate 2.) Russell and Taylor (1937) made similar observations on Mississippi River sands by collecting samples at intervals between Cairo, Illinois, and the Gulf of Mexico which showed no evidence of rounding as a result of attrition during

TABLE 7.—*Minerals in anauxite sands*

From the Sierra Nevada Foothills, Under the Great Valley, and the Coast Range	Andalusite	Tourmaline	Zircon	Epidote	Ilmenite	Leucoxene	Garnet	Biotite	Hornblende	Anauxite	Quartz
Auriferous Channel Muletown	E	D	C	D	D	D				X	X
Ione	E	D	C	D	C	C				X	X
Clements Well at a Depth of 1527 Feet	E	D	E	C	D	D	E		E	X	X
Rio Vista Well at a Depth of 4700 Feet	E	D	D	D	D	D		E	E	X	X
McDonald Is. Well at a Depth of 4950 Feet	E	D	D	D	D	D		E	E	X	X
Potrero Hills	E	D	D	D	D	D	D		E	X	X
Brentwood	E	D	E	C	C	D	D	E	E	X	X
Tesla	E	D	D	D	D	E	E	E	E	X	X

E = rare C = very common
D = common X = present

transportation. Since each of the localities from which anauxite sands were collected is within 30 miles of its nearest neighbor and since some modern sands are angular after traveling that distance, the lack of progressive rounding with increased distance from the source is not a valid argument that some Eocene sands were not transported farther than others. Both the Ione sands and the anauxite sands of the Coast Range are equally angular and could well have been derived from the same source. That this source was the Sierra Nevada can be demonstrated by the similarity in mineral composition of the sands from the Sierra Nevada, under the Great Valley, and in the Coast Range (Table 7). Not only do the samples from these localities possess the same minerals but also the same varieties of blue tourmaline and pink andalusite with certain pleochroism and inclusions which are common throughout the area. Because of the disturbance of these sediments since their deposition, the gradients of the westward-flowing streams transporting

the material cannot be reconstructed accurately. In the foothills belt westward tilting of the Sierra Nevada block has elevated anauxite sediments. At the Clements well anauxite sands lie 1500 feet below the surface and 25 to 30 miles westward at McDonald Island and Rio Vista from 4700 to 5000 feet below the surface. The amount of downwarping due to the weight of the overlying sediments is unknown. Farther westward at Tesla and Brentwood anauxite sediments which were originally at successively lower elevations along the westward-sloping gradients of the Eocene streams now are at the highest elevations because of Coast Range mountain making and tilting. At Tesla, anauxite sands dipping as much as 70 degrees now rise 1500 feet above sea level. Locally, sediments of these Eocene streams have been buried by more than 1 mile of sediment, and gradients have been distorted. Further drilling in the Great Valley may outline more accurately these ancient stream courses.

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St. Louis University, St. Louis, Missouri.

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EXPLANATION OF PLATES

PLATE 1.—PHOTOMICROGRAPHS OF ANAUXITE AND ASSOCIATED SANDS

Figure

(1) Large anauxite plate (A) in anauxite sand at Brentwood, California. x37.

(2) Anauxite plate in sand at Tesla, California. x37.

(3) Pebbly siltstone, which contains type Capay fossils, Capay Valley, California. Q = quartz; F = feldspar. x37.

(4) Anauxite plates in sand at type locality of Ione formation, Ione, California. x37.

(5) Chlorite (C) surrounding anauxite (A) at Silica Company of California Pit, Brentwood, California. x37.

(6) Calcareous fossiliferous sandstone, Silica Company of California Pit, Brentwood, California. Dark grains are ferromagnesian minerals. x37.

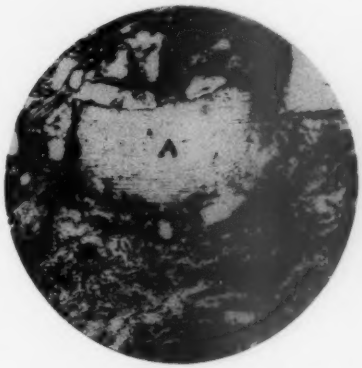
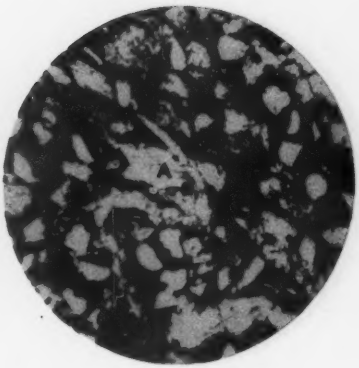


FIGURE 1



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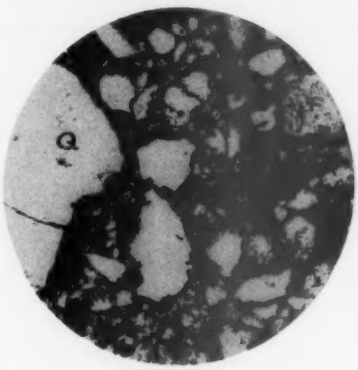
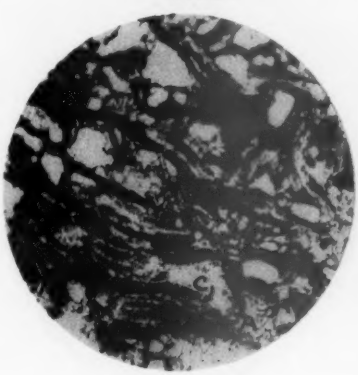


FIGURE 3

FIGURE 4



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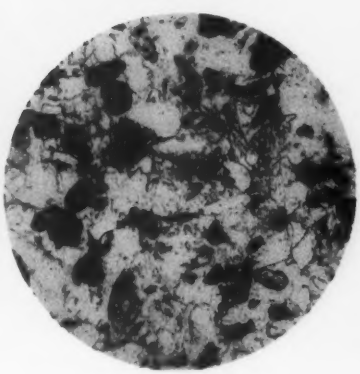
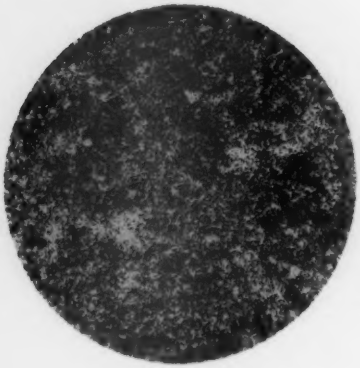


FIGURE 6

PHOTOMICROGRAPHS OF ANAUXITE AND ASSOCIATED SANDS



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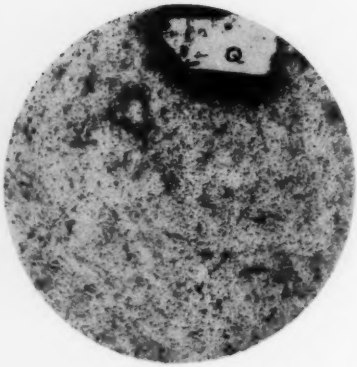

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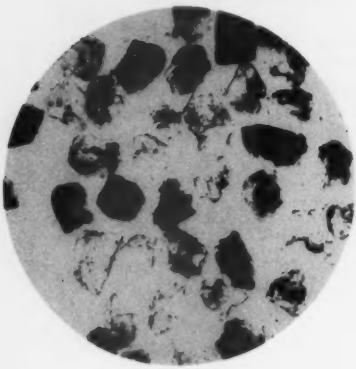

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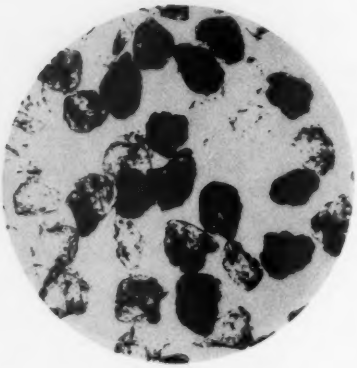

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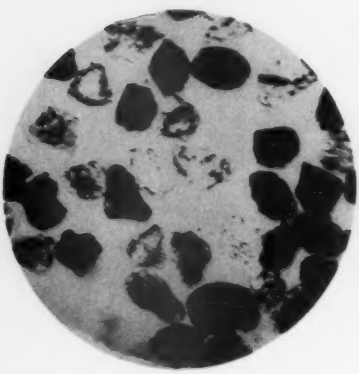

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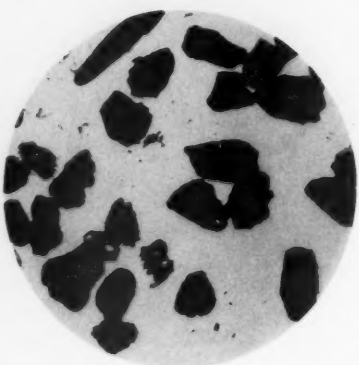

FIGURE 6

PHOTOMICROGRAPHS OF ANAUXITE CLAYS, SANDS, AND RECENT STREAM SANDS

PLATE 2.—PHOTOMICROGRAPHS OF ANAUXITE CLAYS, SANDS, AND RECENT STREAM SANDS

Figure

(1) Anauxite white clay, Clark Pit, Ione. x37.
(2) Anauxite white clay from Tesla Clay Company Mine, Tesla. Q = quartz. x37.
(3) Glass sand, Silica Company of California Mine, Brentwood. x14.
(4) White sand, Tesla. x14.
(5) Recent sand of Calaveras River, above San Andreas, California. x14.
(6) Recent sand of Calaveras River, about 30 miles down stream from locality of sand in Figure 5. x14.